

A Hierarchy of Global Ocean Models Coupled to CESM1

Tien-Yiao Hsu¹, Francois Primeau¹, and Gudrun Magnusdottir¹

¹University of California, Irvine, California, United States

Key Points:

- We develop a flux-corrected globally-coupled ocean-model hierarchy that explicitly includes mixed-layer entrainment and Ekman flow.
- The flux corrections need to contain components stemming from the interannual variability to reproduce the vertical gradient of ocean temperature.
- The hierarchy applied with Arctic sea-ice loss reveals that the ITCZ is sensitive to the entrainment and Ekman flow parameterization.

Abstract

We develop a hierarchy of simplified ocean models for coupled ocean, atmosphere, and sea ice climate simulations using the Community Earth System Model version 1 (CESM1). The hierarchy has four members: a slab ocean model, a mixed-layer model with entrainment and detrainment, an Ekman mixed-layer model, and an ocean general circulation model (OGCM). Flux corrections of heat and salt are applied to the simplified models ensuring that all hierarchy members have the same climatology. We diagnose the needed flux corrections from auxiliary simulations in which we restore the temperature and salinity to the daily climatology obtained from a target CESM1 simulation. The resulting 3-dimensional corrections contain the interannual variability fluxes that maintain the correct vertical gradients of temperature and salinity in the tropics. We find that the inclusion of mixed-layer entrainment and Ekman flow produces sea surface temperature and surface air temperature fields whose means and variances are progressively more similar to those produced by the target CESM1 simulation.

We illustrate the application of the hierarchy to the problem of understanding the response of the climate system to the loss of Arctic sea ice. We find that the shifts in the positions of the mid-latitude westerly jet and of the Inter-tropical Convergence Zone (ITCZ) in response to sea ice loss depend critically on upper ocean processes. Specifically, heat uptake associated with the mixed-layer entrainment influences the shift in the westerly jet and ITCZ. Moreover, the shift of ITCZ is sensitive to the form of Ekman flow parameterization.

Plain Language Summary

Hierarchies of globally coupled ocean models, meaning adding individual ocean processes progressively, provide valuable understandings of the climate system. One of the difficulties is that surface flux corrections are necessary for reproducing a target climate state. In this paper, we manage to overcome this problem and develop a globally coupled ocean model hierarchy that can turn on and off the processes of mixed-layer entrainment and Ekman flow. We use the hierarchy to study the impact of Arctic sea-ice loss on the climate system. We find that the effect of

37 mixed-layer entrainment on ocean heat uptake influences the atmospheric circulation by shift-
38 ing the latitudinal positions of the mid-latitude westerly jet and the Inter-tropical Convergence
39 Zone.

40 **1 Motivation**

41 Model hierarchies contribute to improved understanding of the climate response to exter-
42 nal forcing by isolating the influence of various physical processes (e.g., Claussen et al., 2002;
43 Held, 2005; Jeevanjee et al., 2017; Vallis et al., 2018; Maher et al., 2019). While relatively well-
44 developed hierarchies exist for the atmosphere, less effort has been focused on integrating sim-
45 plified ocean models into global climate models. Inspired by previous regional and intermediate-
46 complexity climate modeling work (Alexander et al., 2000; Haarsma et al., 2005; Alexander &
47 Scott, 2008; Codron, 2012; Hirons et al., 2015), we formulate an ocean-model hierarchy to com-
48 plement the ocean general circulation model (OGCM) in the Community Earth System Model
49 version 1 (CESM1; Hurrell et al., 2013).

50 Simplifications of ocean processes in climate models for studies that focus primarily on the
51 atmosphere typically consist of the following options: (1) prescribing the sea surface tempera-
52 ture (SST) (e.g., Magnusdottir and Saravanan 1999), (2) a single, static layer, known as a slab ocean
53 model (SOM), that can take up heat from the atmosphere, store it, and release it back to the at-
54 mosphere (Kiehl et al., 2006), and (3) a full OGCM (e.g. POP2; R. Smith et al., 2010). Unlike
55 option (1), the SOM accounts for the finite thermal inertia of the upper ocean. The OGCM, in
56 turn, accounts for the three-dimensional transport of heat by ocean currents that respond dynam-
57 ically to surface fluxes of heat, freshwater, and momentum. Because the gap between options (2)
58 and (3) is decidedly large, our goal is to formulate two simplified ocean models that narrow the
59 gap between the SOM and the OGCM.

60 For the first model, we modify the SOM by making the slab thickness a function of time.
61 This time-dependence, which we diagnose from the climatological seasonal cycle in a climate
62 simulation using the OGCM, allows the model to store heat anomalies in the seasonal thermo-

63 cline where they can be re-entrained into the mixed-layer in the following year. The re-emergence
64 of these heat anomalies creates winter-to-winter SST correlations that are particularly important
65 in the North Atlantic and North Pacific Oceans where they are believed to play an important role
66 in the generation of decadal climate variability (Alexander & Deser, 1995; Alexander et al., 1999,
67 2000; Kwon et al., 2011; Newman et al., 2016).

68 In the second model, we add dynamic wind-driven Ekman flows to the previously added
69 time dependence of the mixed-layer depth. The upwelling and downwelling generated by the di-
70 vergence of these flows provide an additional pathway for the atmosphere to interact with the sub-
71 surface ocean. In addition, the Ekman flows contribute to the horizontal transport of heat, which
72 is particularly important in the tropical ocean where it dominates the oceanic heat transport (OHT)
73 (Lee & Marotzke, 1998; Held, 2001). The response of Ekman OHT to changes in surface winds
74 has been shown to be particularly important for damping shifts of the Inter-tropical Convergence
75 Zone (ITCZ) when the Earth is forced with an extra-tropical heat source (Green & Marshall, 2017;
76 Schneider, 2017; Kang, Shin, & Xie, 2018; Kang, Shin, & Codron, 2018; Green et al., 2019).

77 Despite the relative simplicity of our two new models, we need to address several compli-
78 cations: (1) The linear momentum balance equations used to compute the Ekman flows are sin-
79 gular at the equator. To eliminate the singularity we introduce a Rayleigh friction term. Choos-
80 ing an appropriate value for the friction coefficient is the first complication. (2) The simplified
81 model with Ekman flow produces a pronounced equatorial rainfall anomaly that originates from
82 a runaway coupled mode along the equator that needs to be damped using an explicit horizon-
83 tal eddy diffusivity (a mechanistic explanation for this coupled mode is provided in Section 2.6).
84 Choosing an appropriate value for this lateral diffusivity is the second complication that we must
85 address. (3) Finally, because the simplified models neglect many oceanic processes that influ-
86 ence the exchange of heat between the atmosphere and ocean, we need to include prescribed fluxes
87 of heat and moisture in our simplified ocean models to prevent the coupled climate from drift-
88 ing to unrealistic states. Diagnosing these fluxes is the third complication. We address these com-
89 plications in Section 2.

90 In this paper, we illustrate the use of the hierarchy by studying the importance different ocean
91 processes in the response of the climate system to the loss of Arctic sea-ice. Many factors can
92 influence the response, including the background state, wave mean flow interactions in the atmo-
93 sphere, sea-ice physics, and the exchange of heat with the ocean (Screen et al., 2018; D. M. Smith
94 et al., 2022). Of particular importance for motivating our work is the study of Tomas et al. (2016)
95 that identified the importance of oceanic heat transport for modulating the response of the atmo-
96 sphere to sea ice loss by comparing simulations performed with an OGCM to simulations with
97 an SOM.

98 The paper is structured as follows: In Section 2, we introduce the model formulation and
99 solutions to the challenges described above. In Section 3, we investigate the modulating effect
100 of ocean processes on (1) climate variability and (2) the atmospheric response to the loss of Arc-
101 tic sea ice. In Section 4, we present concluding remarks.

102 **2 Methods**

103 **2.1 Hierarchy of ocean models within CESM1**

104 Our hierarchy of climate models is constructed using the Community Earth System Model
105 version 1 (CESM1; Hurrell et al., 2013) by replacing, the standard Parallel Ocean Program ver-
106 sion 2 (POP2, henceforth OGCM; R. Smith et al., 2010; Danabasoglu et al., 2012), in turn with
107 three simplified ocean models. These simplified ocean models, which are generated from a uni-
108 fied numerical code written in Julia (<https://github.com/meteorologytoday/EMOM>), consist
109 of a SOM, a mixed-layer model (MLM), and an Ekman mixed-layer ocean model (EMOM).
110 A Julia interface exchanges the surface fluxes with the Fortran CESM1 coupler.

111 The other components of the climate model consist of the Community Atmosphere Model
112 version 4 (CAM4; Neale et al., 2013), the Los Alamos Sea Ice Model version 4 (Hunke et al., 2010),
113 Community Land Model version 4 (Lawrence et al., 2011), and the River Transport Model (as
114 part of Community Land Model). This configuration of the climate model is equivalent to the
115 Community Climate System Model 4 (CCSM4; Gent et al., 2011).

116 All the simulations are run with nominally $1^\circ \times 1^\circ$ horizontal resolution: The atmosphere
 117 and land models have 192 and 288 points in latitude and longitude on a Gaussian grid (also termed
 118 f09 in CESM1). The ocean and sea-ice models have 384 and 320 points in latitude and longitude
 119 with a displaced-pole grid (also termed g16 in CESM1).

120 2.2 Oceanic general circulation model (OGCM)

121 The Parallel Ocean Program version 2 (POP2) used in our study has a nominal horizon-
 122 tal resolution of about $1^\circ \times 1^\circ$. However, the meridional resolution between 10°S and 10°N —
 123 the range within which ITCZ shifts occur — is better than 0.27° of latitude. Furthermore, the grid’s
 124 “North Pole” is displaced to sit in Greenland, thus avoiding any polar coordinate singularity in
 125 the ocean (R. Smith et al., 2010). In the vertical, the model has 60 levels with separation rang-
 126 ing from 10m in the top 160m of the water column, increasing monotonically from 10m to 250m
 127 in the depth range between 160m, and 3500m, and is fixed at 250m down to the maximum ocean
 128 depth of 5500m. The model’s mixed-layer dynamics are governed by the K-profile parameter-
 129 ization (KPP) vertical-mixing scheme (Large et al., 1994). The model also uses the Gent and McWilliams
 130 (1990) isopycnal mixing scheme.

131 2.3 Ekman mixed-layer ocean model (EMOM)

132 The unified code for our hierarchy of simplified ocean models is called EMOM. We show
 133 the physical processes that are represented by EMOM in schematic form in Figure 1. EMOM is
 134 coupled to the other climate model components through the CESM coupler. EMOM has 33 ver-
 135 tical layers identical to the top 33 layers of the POP2 configuration. These 33 layers cover a to-
 136 tal depth ≈ 503.7 m and range in thickness from 10m to 48m. EMOM solves the following equa-
 137 tions governing the time-evolution of temperature, T , and salinity, S ,

$$138 \frac{\partial T}{\partial t} + \vec{\mathbf{v}}_H \cdot \nabla_H T + w \frac{\partial T}{\partial z} = \frac{\partial}{\partial z} \left(K_V \frac{\partial T}{\partial z} \right) + \nabla_H \cdot (K_H \nabla_H T) - \frac{1}{\rho c_p} \frac{\partial F_T}{\partial z} - \frac{1}{t_R} (T - T_{\text{clim}}) - \frac{\Lambda}{\tau_{\text{FRZ}}} (T - T_{\text{FRZ}}) + \frac{Q_T}{\rho c_p}, \quad (1a)$$

$$139 \frac{\partial S}{\partial t} + \vec{\mathbf{v}}_H \cdot \nabla_H S + w \frac{\partial S}{\partial z} = \frac{\partial}{\partial z} \left(K_V \frac{\partial S}{\partial z} \right) + \nabla_H \cdot (K_H \nabla_H S) - \frac{\partial F_S}{\partial z} - \frac{1}{t_R} (S - S_{\text{clim}}) + Q_S, \quad (1b)$$

141 where $\rho = 1026 \text{ kg/m}^3$ and $c_p = 3996 \text{ J/K/kg}$ are the density and heat capacity of seawater,
 142 F_T is the energy flux consisting of radiation, sensible, and latent heat fluxes, F_S is the virtual salt
 143 flux to account for evaporation minus precipitation, runoff, sea-ice melting and brine injection,
 144 \vec{v}_H is the horizontal velocity, w is the vertical velocity component, h is the mixed-layer thickness,
 145 Q_T and Q_S are heat and salt flux correction terms, ∇_H is the horizontal gradient or divergence
 146 operator, K_H is the horizontal diffusivity. K_H depends on latitude according to

$$147 \quad K_H = K_0 + (K_1 - K_0) \exp\left(-\frac{\phi^2}{2\sigma_K^2}\right) \exp\left(\frac{z}{H_K}\right) \quad (2)$$

148 with ϕ is the latitude in radians, $K_0 = 5.0 \times 10^3 \text{ m}^2/\text{s}$, $K_1 = 2 \times 10^4 \text{ m}^2/\text{s}$, $\sigma_K = 2\pi/36 (= 10^\circ)$
 149 and $H_K = 100 \text{ m}$. The value of K_0 is referenced from Nummelin et al. (2021) in which the SST
 150 diffusivity is diagnosed (see their Figure 2). The values of K_1 and σ_K are chosen to prevent the
 151 spurious rainfall pattern mentioned in the introduction and further discussed in Section 2.6. K_V
 152 is the vertical diffusivity that depends on the mixed-layer thickness and vertical gradient of buoy-
 153 ancy, b . Explicitly,

$$154 \quad K_V = \begin{cases} 1 \text{ m}^2/\text{s} & \text{if } -h < z < 0 \text{ or } \frac{\partial b}{\partial z} < 0, \\ 1 \times 10^{-4} \text{ m}^2/\text{s}, & \text{otherwise.} \end{cases} \quad (3)$$

155 The short-wave radiative heating is divided into two components and each follows an exponen-
 156 tial depth profile with a constant e -folding depth. The parameters are selected according to Type
 157 I water in Table 2 of Paulson and Simpson (1977). The remaining surface heat fluxes and virtual
 158 salt fluxes are implemented as interior sources in the top 10-meter-thick layer of the model so that
 159 Equations (1a) and (1b) can be solved subject to no-flux boundary conditions.

160 To compute the buoyancy, we approximate the density using a third-order polynomial ap-
 161 proximation for the equation of state as described in Bryan and Cox (1972). In the numerical im-
 162 plementation, K_V is specified at the interface between the grid boxes. The effective mixed-layer
 163 thickness must therefore coincide with the discrete depths of the interfaces between the grid boxes.
 164 T_{clim} and S_{clim} are the 50-year reference monthly climatological profiles that the entire ocean is
 165 weakly restored to with a timescale $t_R = 100$ years. With this choice of t_R , we refer to this term
 166 as the “weak-restoring” term. To take into account the latent heat exchanges associated with the

167 transformation of water between the liquid and solid phases, we introduce the parameters $\tau_{\text{FRZ}} =$
 168 1 day and $T_{\text{FRZ}} = -1.8^\circ\text{C}$ as the freezing timescale and freezing point of ocean water. Λ is the
 169 Heavyside step function, i.e.,

$$170 \quad \Lambda = \begin{cases} 1 & \text{if } T < T_{\text{FRZ}}, \\ 0 & \text{otherwise.} \end{cases}$$

171 This design switches on the freezing process so that ocean temperature is forced to be above the
 172 freezing point. The latent heat released is sent to the coupler so that sea ice may grow. For sim-
 173 plicity, we prescribe the space and time dependence of h . We considered adopting a dynamic mix-
 174 ing layer model such as the KPP scheme (Large et al., 1994) as used in the POP2 model or a Niiler-
 175 Kraus type bulk formulation (Niiler, 1977; Gaspar, 1988). However, the former needs as input
 176 the vertical shear, which is not available in our simplified model, and the latter, which we imple-
 177 mented and tested, produces a mean mixed-layer thickness that is substantially different than the
 178 one in the parent simulation. This difference in the mean state complicates the interpretation of
 179 our experiments. For these reasons, we focus here on the simplified models with a prescribed time-
 180 dependence for the mixed-layer thickness.

181 As for the velocity field, we use the the same formulation as Codron (2012) in which the
 182 Ekman velocity is diagnosed from the instantaneous wind-stress using a linear momentum bal-
 183 ance:

$$184 \quad \vec{v}_{\text{EK}} = \frac{1}{\rho H_{\text{EK}}(\epsilon^2 + f^2)} (f\tau^y + \epsilon\tau^x, -f\tau^x + \epsilon\tau^y) \quad (4)$$

185 where $H_{\text{EK}} = 50\text{m}$ is the Ekman layer thickness, f the latitude-dependent Coriolis parameter,
 186 and $\vec{\tau} = (\tau^x, \tau^y)$ is the surface wind stress, ϵ the Rayleigh friction coefficient that removes the
 187 singularity of Ekman solution at the equator. The value of ϵ is set to $\epsilon = 1.4 \times 10^{-5} \text{ s}^{-1}$ (see Sec-
 188 tion 2.4 for detail) so that the Rayleigh friction dominates over the Coriolis force in a narrow band
 189 between 5.5°S and 5.5°N . At the equator \vec{v}_{EK} becomes aligned with the direction of surface wind
 190 stress. We will call the part of \vec{v}_{EK} that is proportional to $\epsilon\vec{\tau}$ the frictional Ekman flow.

191 Two alternative parameterizations have been proposed for removing equatorial Ekman-balance
 192 singularity. In terms of the cross-equatorial ocean heat transport, these parameterizations pro-
 193 duce a better agreement between the OGCM and the simplified ocean model. However, there are

194 other problems with these alternative parameterizations. The parameterization of Kang, Shin,
 195 and Codron (2018) drops the ϵ term in the numerator of the second component of Equation (4).
 196 Unfortunately, this simplification cannot be justified from the original momentum equation. In
 197 order to drop the ϵ dependence in the numerator of the meridional component of Equation (4),
 198 the Rayleigh friction term in the zonal momentum equation must vanish, but without it, the zonal
 199 component of Equation (4) would still be singular at the equator. The other parameterization (Afargan-
 200 Gerstman & Adam, 2020) was proposed in the context of an aqua-planet simulation where the
 201 zonal symmetry causes the mean meridional wind component to vanish so that only the merid-
 202 ional Ekman flow needs to be considered. For this special case, the Ekman-balance singularity
 203 can be eliminated by retaining the latitudinal-dependence of the Coriolis parameter, i.e. the β ef-
 204 fect, and noting that in the absence of friction the meridional Ekman flow is proportional to the
 205 curl of the wind stress divided by β . No such balance is available for the zonal Ekman flow, which
 206 will still be singular if it is present. Because we are considering a realistic continental configu-
 207 ration, we decided to adopt Codron's original formulation, i.e. Equation (4).

208 As in Codron (2012), we put the return flow directly below the Ekman layer with a volume
 209 flux in the opposite direction and of equal magnitude to the surface Ekman flow. That is,

$$210 \quad \vec{v}_{\text{RF}} = -\frac{H_{\text{EK}}}{H_{\text{RF}}} \vec{v}_{\text{EK}}, \quad (5)$$

211 where $H_{\text{RF}} \approx 453.7$ m is the thickness of the return flow so that the resulting thickness conve-
 212 niently coincides with OGCM grid. Thus the horizontal velocity is \vec{v}_{EK} $z \in (-H_{\text{EK}}, 0]$ and \vec{v}_{RF}
 213 for $z \in (-(H_{\text{EK}} + H_{\text{RF}}), -H_{\text{EK}}]$. This parameterization ensures that the vertically integrated \vec{v}_{EK}
 214 is zero.

215 The resulting Ekman OHT is easy to diagnose given the temperature profile in each wa-
 216 ter column. For any given water column, Ekman OHT is given by

$$217 \quad \begin{aligned} \text{OHT}_{\text{EK}} &= \rho_0 c_p \int_{z=-(H_{\text{EK}}+H_{\text{RF}})}^0 v_{\text{EK}} T dz \\ 218 \quad &= \rho_0 c_p (H_{\text{EK}} v_{\text{EK}} T_{\text{EK}} + H_{\text{RF}} v_{\text{RF}} T_{\text{RF}}) \\ 219 \quad &= c_p \Delta T \left(\frac{-f\tau^x + \epsilon\tau^y}{\epsilon^2 + f^2} \right) \end{aligned} \quad (6)$$

220

221 where

$$222 \quad T_{\text{EK}} = \frac{1}{H_{\text{EK}}} \int_{-H_{\text{EK}}}^0 T \, dz, \text{ and } T_{\text{RF}} = \frac{1}{H_{\text{RF}}} \int_{-(H_{\text{EK}}+H_{\text{RF}})}^{-H_{\text{EK}}} T \, dz \quad (7)$$

223 are the mean temperatures over the Ekman and return-flow layers, and

$$224 \quad \Delta T \equiv T_{\text{EK}} - T_{\text{RF}} \quad (8)$$

225 is the difference between the mean temperatures defined above. In going from the second to the
 226 third line of Equation (6), we used Equation (5). Given typical tropical wind stress $(\tau^x, \tau^y) =$
 227 $(0.01, 0)\text{N/m}^2$ at latitude 5.5°N where $\epsilon = f$ (Section 2.4 shows how we pick ϵ), the computed
 228 zonally integrated $(2\pi \cos 5.5^\circ \times (\text{Radius of earth}) \approx 4 \times 10^7\text{m})$ meridional OHT is about 0.06
 229 PW per $^\circ\text{C}$ change of ΔT .

230 On the numerical side, EMOM is discretized using Arakawa C-grid. The model exchanges
 231 surface fluxes with the atmosphere every 24 hours. The time step is set to three hours. Advec-
 232 tion and horizontal diffusion are implemented using the QUICKEST scheme (upwind, second-
 233 order accuracy; Leonard, 1979). Finally, the vertical diffusion, weak-storing, and freezing pro-
 234 cess are solved using the implicit Euler backward method.

235 All the members of our hierarchy except for the OGCM can be recovered from the same
 236 code base. To recover the MLM, we simply set $\vec{\mathbf{v}} = 0$. To recover the SOM, we (i) set $\vec{\mathbf{v}}_H = \mathbf{0}$,
 237 (ii) set the mixed-layer thickness of each grid point to its time-averaged value, and (iii) replace
 238 the topography mask to match the time-averaged mixed-layer thickness. With these settings, $w_e =$
 239 0 and $K_v = 1 \text{ m}^2/\text{s}$ everywhere, which makes our model equivalent to a well-mixed slab.

240 There are two major differences between CESM1-SOM and our SOM. First, CESM1-SOM
 241 sets the horizontal diffusivity K_H to zero while our SOM has non-zero spatially varying K_H given
 242 by Equation (2). This choice simplifies the task of comparing the output of the different mod-
 243 els of our hierarchy. Second, the CESM1-SOM uses the boundary layer depth output from POP2
 244 (variable HBLT) instead of mixed-layer depth (variable HMXL), which we use in EMOM, for the
 245 mixed-layer thickness. The boundary layer depth and the mixed-layer thickness are equal when
 246 they are at their maximum, but otherwise the boundary layer is thinner than the mixed-layer. Past

247 studies have shown that a shallower annual-mean mixed-layer thickness leads to colder SSTs, which
 248 tends to increase the albedo through cloud-radiative feedback processes (Donohoe et al., 2014;
 249 Wang et al., 2019). However, the flux corrections that are applied to our simplified models will
 250 generally eliminate this effect on the mean climate. The choice of thickness would most likely
 251 also affect the climate variability, but the above mentioned studies do not offer any specific rec-
 252 ommendation as to which choice is preferable in a SOM. We, therefore, chose to set h to HMXL.

253 The full OGCM completes our hierarchy as the most realistic ocean model. Thus, the re-
 254 sulting hierarchy of ocean models from simplest to most complex are: the SOM, the MLM, the
 255 EMOM, and the OGCM. As we progress up the hierarchy the MLM includes entrainment, which
 256 is absent in the SOM, the EMOM includes Ekman transport, which is absent in the MLM and
 257 the OGCM includes everything else (i.e. gyres, overturning circulation, waves, etc.)

258 **2.4 Estimation of Rayleigh Friction ϵ in Equation (4)**

259 To avoid the singularity of the Ekman flow solution at the equator, we introduced a Rayleigh
 260 friction term in the horizontal momentum equations used to derive Equation (4). To determined
 261 the value of ϵ , we minimized the sum of the squared differences between the vertical velocities
 262 computed by the OGCM and those computed from the divergence of $\vec{v}_{EK}(\epsilon)$ integrated from the
 263 surface, assumed flat and rigid, down to a depth of $H = 50$ m in the latitude band between 10°S –
 264 10°N . Explicitly,

$$265 \quad \epsilon = \underset{\epsilon}{\operatorname{argmin}} \sum_{i \in \mathcal{X}} |w_{\text{OGCM}}^i - w_{\text{EK}}^i(\epsilon)|^2 \quad (9)$$

266 where

$$267 \quad w_{\text{EK}}^i(\epsilon) = H \nabla \cdot \vec{v}_{\text{EK}}(\epsilon), \quad (10)$$

268 and \mathcal{X} denotes the set of mesh points between 10°S and 10°N .

269 The solution to the above non-linear least-squares problem yielded a value of $\epsilon = 1.4 \times 10^{-5} \text{ s}^{-1}$.
 270 With this choice for ϵ , Rayleigh friction dominates over the Coriolis force between 5.5°S – 5.5°N .
 271 Thus within 5° of the equator, the Ekman flow is more parallel than perpendicular to the direc-
 272 tion of the wind stress. At the equator, the Ekman flow is exactly aligned with the wind stress.

2.5 Derivation of correction flux

For the simulations that use SOM, MLM, and EMOM, we apply flux corrections to force the simulated climates towards the one simulated using CESM1 (i.e. CTL_OGCM; see Section 2.7). The flux corrections make up for the mean heat and freshwater transports captured by the full OGCM but not explicitly represented in our simplified models. These flux corrections include rectification effects due to interannual variability. For example, if we decompose the variables in Equation (1a) into a mean annual cycle plus anomaly, i.e. $(\cdot) = \overline{(\cdot)} + (\cdot)'$, and then apply the averaging operator $\overline{(\cdot)}$, we obtain

$$\begin{aligned} \frac{\partial \bar{T}}{\partial t} + \bar{\mathbf{v}}_{\text{EK}} \cdot \nabla_H \bar{T} + \bar{w}_{\text{EK}} \frac{\partial \bar{T}}{\partial z} = \frac{\partial}{\partial z} \left(K_V \frac{\partial \bar{T}}{\partial z} \right) + \nabla_H \cdot \left(K_H \nabla_H \bar{T} \right) - \frac{1}{\rho c_p} \frac{\partial \bar{F}_T}{\partial z} - \frac{1}{t_R} \left(\bar{T} - T_{\text{clim}} \right) \\ - \frac{\bar{\Lambda}}{\tau_{\text{FRZ}}} \left(\bar{T} - T_{\text{FRZ}} \right) + \left[\frac{\bar{Q}_T}{\rho c_p} - \overline{\mathbf{v}'_{\text{EK}} \cdot \nabla_H T'} - \overline{w'_{\text{EK}} \frac{\partial T'}{\partial z}} + K'_V \frac{\partial^2 T'}{\partial z^2} - \frac{\Lambda' T'}{\tau_{\text{FRZ}}} \right]. \end{aligned} \quad (11)$$

The terms in the square bracket correspond the flux corrections needed to reproduce the mean annual cycle. In addition to the effect of missing processes, \bar{Q}_T , the bracket includes anomaly-product terms. Some of the interannual variability that produces the anomalies that contribute to these product terms will be generated spontaneously by the simplified models. Their effects do not need to be included in the flux correction. However, some of the interannual anomalies originate from processes that can only be simulated using the full OGCM and will, therefore, be absent in our simplified climate models. As a prominent example, the El Niño-Southern Oscillation (ENSO), which dominates in the tropics, cannot be captured in our simplified models. Its effect on the annual mean fluxes must be included as part of the flux correction. Without it, our models cannot reproduce the correct vertical temperature gradient. There is, however, a potential risk of double-counting the effect of anomalies that can simulated in the simplified coupled climate models.

To derive the flux correction given by the terms in the square bracket of Equation (11), we first use the 50-year daily climatology of temperature and salinity (T_{clim} and S_{clim}) from the end of a 1000-year spin-up run. Then, for each model, we temporarily set $t_R = 15$ days in the simplified model and run the resulting coupled climate model for 20 years. We then record the monthly

300 mean values of the restoring term for the last 19 years. Finally, we average each month to pro-
 301 duce a three-dimensional flux correction that is periodic with a period of one year.

302 **2.6 Importance of large K_H near the equator**

303 Without explicit horizontal diffusion, EMOM produces a persistent band of excessively strong
 304 equatorial precipitation in the Central Pacific (Figure S1a, 150-170°W with the zonal mean shown
 305 in Figure S1b).

306 In a narrow band along the equator where the frictional force dominates over the Corio-
 307 lis, convergent surface winds (Figure S1b, red dashed line) above a warm SST anomaly (Figure
 308 S1c, black solid line) also drive convergent ocean currents that suppress upwelling (Figure S1c,
 309 red dashed line). Off the equator, where the Coriolis force dominates over the friction force the
 310 wind-driven currents are divergent and drive upwelling. Further examination of the meridional
 311 wind stress τ^y and ocean 50-m vertical velocity w_{50m} reveals that this is a consequence of the fol-
 312 lowing coupled positive feedback. Since our Rayleigh friction dominates over the Coriolis force
 313 within 5.5° of the equator, anomalous wind convergence drive anomalous SST convergence near
 314 the equator. The resulting warming of the sea surface acts to reinforce the convection in the over-
 315 lying atmosphere. Furthermore, the upwelling of cold waters in regions straddling the center of
 316 atmospheric convection induces locally descending motion in the atmosphere, which acts to fur-
 317 ther reinforce the the convergent winds and atmospheric convection.

318 In the real ocean, the generation of the strong temperature gradients seen in Figure S1c would
 319 lead to a baroclinic instability that would generate a vigorous eddy field. The diffusive fluxes gen-
 320 erated by these eddies would come to balance the advective flux due to the mean flow. By posit-
 321 ing a balance between the advective and diffusive fluxes, we can estimate an appropriate value
 322 of K_H so that we can suppress the spurious rain band in EMOM,

$$323 \quad K_H \frac{\partial^2 T}{\partial y^2} = -v \frac{\partial T}{\partial y} \quad (12)$$

$$324 \quad \Rightarrow K_H \approx |v \Delta y| \quad (13)$$

325

326 where Δy is the meridional length scale. From Figure S1 we estimate that $\Delta y \approx 400$ km, and that
 327 the maximum meridional ocean Ekman flow is about $\max(|\tau^y|)/(\rho_0 \epsilon 50 \text{ m}) \approx 0.028$ m/s. From
 328 these scales, we estimate $K_H = |v \Delta y| = 1.12 \times 10^4$ m²/s. Thus, to ensure the suppression of cou-
 329 pled positive feedback, we set K_1 in Equation (2) to be 2×10^4 m²/s.

330 2.7 Simulation design

331 Since we want to test how ocean processes modulate climate variability and the response
 332 to Arctic sea-ice loss, it is necessary to have a set of control runs where all hierarchy members
 333 produce the same climate.

334 The control runs are initialized with the January-mean fields from the end of the 1000-year
 335 spin-up simulation with the full OGCM. They are performed with the necessary correction fluxes
 336 derived in Section 2.5 and are labeled as CTL_[MODEL] where [MODEL] can be one of SOM,
 337 MLM, EMOM, or OGCM. The CTL simulations are run for 120 years and the last 100 years are
 338 analyzed. The biases of each model are defined as the deviations from the CTL_OGCM simu-
 339 lation as opposed to the observed mean state.

340 The perturbation runs are performed with projected Arctic sea-ice loss such that sea-ice
 341 thickness matches the mean sea-ice thickness of the year 2081–2100 of RCP8.5 simulation. The
 342 details of how the forcing is applied are explained in Appendix A. As we did for the CTL runs,
 343 the sea-ice-loss runs are denoted by SIL_[MODEL]. Also, we denote the entire set of sea-ice-
 344 loss runs as SIL and define $\text{RESP_}[MODEL] := \text{SIL_}[MODEL] - \text{CTL_}[MODEL]$.

345 SIL is run for 180 years. SIL_SOM quickly reaches equilibrium within 50 years whereas
 346 the rest of the models adjust quickly in the first 80 years and slowly drift afterward (Figure S2a).
 347 In SIL_OGCM, the AMOC weakens during the first 30 years followed by a recovery during the
 348 next 50 years, and then stabilizes (Figure S2b). Therefore, we analyze years 81–180 and use this
 349 period to represent the decadal and centennial adjustment of the climate system.

3 Results

3.1 Assessment of the mean states

This section presents the analysis of the CTL runs. The experimental setup is documented in Section 2.7.

The simulated climates agree more with the CTL_OGCM as we include more ocean processes. We here discuss the sea-ice and SST biases as they are two important surface properties that control the climate.

3.1.1 *Sea-ice area is better constrained than sea-ice volume*

For each CTL simulation with a simplified ocean model, we compared the simulated sea-ice area and the sea-ice volume to the area and volume simulated using the OGCM. The results are summarized in Table 1. The relative sea-ice area biases range from -15% to +2%, while the relative sea-ice volume bias is -30% to +3%. We find a smaller bias for the area compared to the volume in both hemispheres. This larger bias for the volume compared to the area is due to latent heat fluxes in the presence of sea ice that are invisible to our method for diagnosing the flux corrections – recall that our flux corrections are diagnosed from a restoring term that acts on the water temperature (Section 2.5).

Another possible reason for sea-ice biases is the double-counting of the $\overline{\Lambda'T'}$ term in Equation (11), as discussed in Section 2.5. In high latitudes, an important role of the ocean is to serve as a heat reservoir, meaning that the SOM can produce a portion of the interannual variability. Since lower ocean temperature activates the freezing, Λ' and T' are negatively correlated, i.e., $\overline{\Lambda'T'} < 0$. Therefore, if the term $\overline{\Lambda'T'}$ is a significant contributor to the flux correction and a significant fraction of $\overline{\Lambda'T'}$ in the CTL_OGCM, including the associated heat fluxes contributes to the warm SST biases we see in Figure 2.

3.1.2 *Entrainment reduces SST bias*

In all simplified models, the SST biases in most of the regions are within 0.5°C (Figure 2). The biases in the CTL_SOM have a similar pattern to those obtained using the CCSM3-SOM (c.f. Figure 2 of Bitz et al., 2012): warmer SSTs over the tropical Eastern Pacific and the Southern Ocean and colder SSTs along the Kuroshio Extension. Once the mixed-layer dynamics are included, these biases are significantly reduced. The tropical warm bias in CTL_SOM causes more precipitation than CTL_MLM and CTL_EMOM (Figure 3).

The common warm SST bias regions are along the sea-ice edge in the Southern Ocean, with CTL_EMOM having the largest bias. This contradicts our expectation that SST bias should decrease as more ocean processes are included. Since the location of this warm SST bias is along the sea-ice edge, it might be due to the double-counting of $\overline{\Lambda'T'}$ the same issue as discussed in the previous section on sea-ice bias.

3.2 *The impact of ocean processes on variability*

Even though hierarchy members produce a similar climate mean states, their variability may differ. Here, we define “variability” as the standard deviation of the anomaly time series of the given variable. The anomaly is the deviation from its climatology (monthly, seasonally, or annually depending on the context).

We also examine the “re-emergence” of SST anomaly (SSTA). It refers to the 12-month lag correlation of SSTA in winter (Alexander et al., 2000). This memory effect is a consequence of seasonal mixed-layer entrainment, which leads to enhanced predictability.

3.2.1 *Variability of SST and surface air temperature is improved*

Because our simplified models cannot generate most of the tropical variability due to the lack of Kelvin and Rossby waves dynamics, direct comparison between full OGCM with the simplified models would be harder to interpret. Thus, we define CTL_OGCM* as the filtered CTL_OGCM

397 in which we subtract out the projection of the variability on the first empirical orthogonal func-
398 tion of the monthly SSTA between 20°S–20°N.

399 In general, the SST variability is similar to that of CTL_OGCM*, with higher variability
400 along the storm track (Figure 4). We also notice that SST variability decreases by about 10–30%
401 in CTL_MLM and CTL_EMOM compared to CTL_SOM (Figure 4) even though they have sim-
402 ilar atmospheric variability (sea-level pressure, not shown). Since CTL_SOM does not have mixed-
403 layer entrainment, it highlights the importance of the seasonal entrainment in winter that acts to
404 dampen the stochastic atmospheric forcing. We further verify that mixed-layer entrainment can
405 effectively reduce the SST variability by constructing a stochastic one-dimensional mixed-layer
406 model in Appendix B.

407 Once we include the Ekman flow, the shape of the SST variability pattern in the simplified
408 model of certain regions resembles more to that in the OGCM (Figure 4). These regions are North-
409 ern Pacific (box A), Northern Atlantic (box B), and Pacific-Atlantic sector of Southern Ocean (box
410 C). In particular, the tongue-shaped SST variability pattern extending eastward from Japan (box
411 A) in CTL_EMOM resembles the pattern of Pacific Decadal Oscillation (PDO) SST variability.
412 Since we cannot observe this in CTL_SOM and CTL_MLM, it shows that Ekman is an essen-
413 tial process reshaping SST variability pattern. This is consistent with Newman et al. (2016) that
414 a portion of PDO SST variability originates from coupled effect of the weather noises with Ek-
415 man flow. Our hierarchy conveniently demonstrate this effect. Besides, the magnitude of the SST
416 variability in box A is not as strong as the variability in CTL_OGCM*, implying that the Kuroshio
417 Extension variability, which is not captured in our simplified models, is an important contribu-
418 tor of SST variability in the North Pacific. Similarly, the reduced SST variability bias is accom-
419 panied by a reduction in winter surface-air-temperature variability. In the Southern Ocean (Fig-
420 ure. 5 box A), both mixed-layer entrainment and Ekman flow reduce the biases. In North Amer-
421 ica (box B), including Ekman flow reduces the bias, likely because it locates downstream of the
422 regions where SST variability biases are reduced.

423 **3.2.2 SSTA re-emergence improvement**

424 In CTL_OGCM, we find stronger re-emergence signals in higher latitudes during winter,
425 specifically in the North Pacific (Figure 6 box A) and Southern Ocean (Figure 6 box B). While
426 the re-emergence bias is reduced in the North Pacific (box A), it is too strong in the Southern Ocean
427 (box B). We speculate that in the Southern Ocean, the northward transport of the subducted Antarc-
428 tic Intermediate Water and diffusive effect of eddies remove much of the signal that leads to low
429 winter-to-winter correlation in the CTL_OGCM. Without these processes, winter-to-winter cor-
430 relation will be too strong.

431 In all simplified models, a strong winter-to-winter correlation is present in the tropical East-
432 ern Pacific while there is none in CTL_OGCM (Figure 6 box C). Although we are unsure of the
433 causes to the low winter-to-winter correlation in CTL_OGCM, we speculate that it is due to the
434 presence of the horizontal currents and wave dynamics that might act to damp the temperature
435 anomalies.

436 **3.3 Response to sea-ice loss**

437 This section presents the analysis of the response of climate to Arctic sea-ice loss during
438 year 81–180. The experiment detail is documented in Section 2.7.

439 **3.3.1 Roles of entrainment and AMOC slowdown in ocean heat uptake**

440 All the models in the hierarchy show a significant SST warming in the northern hemisphere
441 due to Arctic sea-ice loss forcing (Figure 7), whereas the southward extent of the warming gets
442 weaker as more ocean processes are included. In the North Pacific, the southward extent of the
443 warming in RESP_EMOM is similar to RESP_OGCM, especially the warm SST tongue in the
444 center. In contrast, in the North Atlantic the decrease of SST in RESP_OGCM, also known as
445 the “SST warming hole”, is not simulated in any of the simplified models. The warming hole in
446 RESP_OGCM seems to reduce the warming extent over the North Pacific as well.

447 The warming extent of SST in SOM is the largest, which is consistent with the fact that the
448 SOM traps the heat in the ocean surface layer. Furthermore, including mixed-layer entrainment
449 and Ekman flow is insufficient to reproduce the reduced extent of warming in the North Atlantic
450 of RESP_OGCM. The SST warming is better captured in the North Pacific than in the North At-
451 lantic in our simplified models. Since AMOC responds strongly to sea-ice loss (Figure S2b), we
452 speculate that AMOC in the OGCM efficiently removes the anomalous heat after being transferred
453 to the subsurface in the Atlantic. The role of the AMOC slowdown can be understood from the
454 rough alignment of the anomalous AMOC upwelling (Figure 8) with the latitudinal location of
455 the SST warming hole in RESP_OGCM (Figure 7) between 40°N–60°N. The slowdown of AMOC
456 creates an anomalous heat flux divergence between 40°N–60°N that removes the anomalous heat
457 entrained from the surface. Thus, the efficient ocean heat uptake in RESP_OGCM is a combined
458 effect of entrainment and AMOC slowdown.

459 ***3.3.2 The response of the position of westerly jet is sensitive to ocean heat uptake***

460 The atmospheric zonal mean temperature response to Arctic sea-ice loss aligns latitudi-
461 nally with the warming response in the ocean in all the models (Figure 9). The horizontal and
462 vertical extent of the warming decreases as we include more ocean processes. Since the response
463 follows the thermal wind relation (not shown), we can use the resulting different meridional gra-
464 dients of the zonal mean temperature to explain the mean jet response. For example, we see that
465 in RESP_SOM, the atmospheric warming is so extensive that it reduces the meridional temper-
466 ature gradients on both sides of the westerly jet, resulting in an overall weakening of the jet. The
467 gradients in RESP_MLM and RESP_EMOM over the northern edge of the jet decrease much
468 more than to the south, which weakens the westerly jet and shifts it southward. Furthermore, in
469 RESP_OGCM the temperature gradient is decreased to the north but increased to the south such
470 that the jet is shifted rather than weakened. We conclude that the response of the westerly jet to
471 sea-ice loss is sensitive to ocean processes that take up the heat.

3.3.3 *The response of ITCZ position is sensitive to surface ocean processes*

The response of tropical precipitation to Arctic sea-ice loss highlights the connection between sea-ice loss and ITCZ shifts (Figure 10). The ITCZ shifts northward, into the warmed hemisphere, in all models but with various extents and shapes. In RESP_SOM, there is a significant northward shift of ITCZ across the globe and this signal is reduced in RESP_MLM and amplified in RESP_EMOM. Note that RESP_OGCM shows a moderate ITCZ shift in the Atlantic Ocean and mostly a narrowing of the rainband towards the equator in other basins.

The shift in the ITCZ is a consequence of the excessive energy that is injected into the Arctic and transported across the equator into the southern hemisphere. This anomalous planetary heat transport (PHT) is partitioned into atmosphere heat transport (AHT) and OHT. The ITCZ shift is correlated with the anomalous AHT (Schneider et al., 2014). Here, we expect that the ITCZ shift may be suppressed in two ways: by (1) ocean heat uptake, or (2) southward cross-equatorial OHT through Ekman flow.

The first case, ocean heat uptake, applies to RESP_MLM and RESP_OGCM. Comparing the zonal mean of precipitation response of RESP_MLM to RESP_SOM (Figure 11), we see that RESP_MLM and RESP_SOM peak at the same latitude on the northern side but RESP_MLM is only half the strength. This is because the heat is sent to the subsurface through the entrainment and temporarily reduces the imbalance of inter-hemispheric energy budget. Since RESP_OGCM takes up the heat even more efficiently with the aid of anomalous AMOC, its ITCZ shift is the weakest (Figure 10).

The second case, Ekman modulation, stems from the physical argument that the response of zonal wind stress in both hemispheres drives a southward Ekman OHT (Schneider, 2017; Green & Marshall, 2017; Green et al., 2019; Kang, 2020; Adam, 2021). However, our simulation RESP_EMOM shows the opposite. While weakened trade winds in the northern hemisphere and strengthened trade winds in the southern hemisphere (Figure 12 solid line) should produce southward Ekman OHT, the northward wind stress on the equator (dashed line) generates a northward OHT. The

498 net result of these two regimes drives heat convergence (divergence) in the northern (southern)
 499 hemisphere resulting in amplified northward ITCZ shift.

500 To understand the apparent inconsistency between the amplified ITCZ shift in RESP_EMOM
 501 and the expectation from past studies arguing that Ekman flow should damp the shift, we exam-
 502 ine the Ekman OHT, i.e. OHT_{EK} defined in Equation (6). OHT_{EK} has two contributions, one from
 503 $-f\tau^x$ and the other from $\epsilon\tau^y$. Past studies omitted the contribution from $\epsilon\tau^y$. However, there
 504 is evidence from OGCM simulations for the existence of the frictionally-driven overturning. For
 505 example, Figure 9 of Jayne and Marotzke (2001) shows a shallow frictionally driven Ekman over-
 506 turning cell within the top 100m of the ocean surface. Also, a shallow clockwise overturning cell
 507 near the surface at the equator that is superimposed onto the deeper and stronger counterclock-
 508 wise density-driven overturning cell is also shown in Figure 5c and 5f of Green et al. (2019). Al-
 509 though the frictionally-driven cell is poorly resolved, its effect on the OHT is visible in Figure
 510 9b of Green et al. (2019). The bump in the OHT-v.s.-latitude plot at the equator is a manifesta-
 511 tion of the anomalous frictionally-driven Ekman heat transport. However, Equation (6) also shows
 512 that OHT_{EK} is proportional to ΔT , which can be directly modified by the choice of $H_{total} = H_{EK} +$
 513 H_{RF} , the total thickness for the Ekman and return flows. As a result, our simplified model could
 514 potentially sensitive to the choices of H_{EK} and H_{RF} .

515 We summarize this discussion in Figure 13 that shows the decomposition of the heat trans-
 516 port in each model. In the RESP_SOM and RESP_MLM, almost all the response is through AHT
 517 with a small contribution to OHT. The contribution to OHT due to diffusion is small. The AHT
 518 in RESP_MLM is smaller than for RESP_SOM because the ocean takes up the heat. In RESP_OGCM,
 519 the ocean takes up heat and also transports it horizontally so that the subsurface cold water can
 520 be replenished to efficiently take up the heat. In RESP_EMOM, the OHT at the equator is north-
 521 ward due to frictional Ekman flow, resulting in an amplification of the ITCZ shift.

522 **4 Conclusion**

523 In this paper, we constructed a hierarchy of ocean models including SOM, MLM, EMOM,
524 and OGCM (POP2) that separates mixed-layer entrainment and Ekman flow by progressively in-
525 cluding each process into the model. We further couple it to the climate model CESM1 with re-
526 alistic topography and successfully derive three-dimensional correction flux through the nudg-
527 ing method so that the appropriate climatologies for SST, precipitation, and sea-ice area are re-
528 produced.

529 The mixed-layer entrainment is important for multiple reasons. In the control climate, we
530 found that mixed-layer entrainment significantly reduces the SST bias in SOM. The entrainment
531 also reduces SST variability that is exaggerated in SOM, since it introduces cold deep water into
532 the mixed-layer during winter when interannual variability is the strongest. Entrainment also re-
533 sults in stronger winter-to-winter SSTA correlation in agreement with the OGCM almost every-
534 where. There are, however, two prominent exceptions. In the Southern Ocean and tropical East-
535 ern Pacific, the simplified models produce too high winter-to-winter SSTA correlations. Because
536 horizontal transport would remove local temperature anomalies, we speculate that the high cor-
537 relation is due to their absence in the simplified models. In the sea-ice loss perturbation exper-
538 iment, including entrainment allows the deep ocean to take up the excessive heat associated with
539 sea-ice loss and reduces the spatial extent of the warming in the atmosphere. Thermal wind re-
540 lation then translates the anomalous atmospheric warming structure into a latitudinal shift of the
541 westerly jet.

542 Moreover, the fact that mixed-layer entrainment alone cannot efficiently remove the heat
543 by horizontal transport highlights the function of AMOC: the AMOC removes the subsurface wa-
544 ter carrying the heat entrained from the surface and replaces it with cold water. The uptake of
545 thermal energy also dampens the ITCZ shift because it temporarily reduces the inter-hemispheric
546 imbalance in the heat budget. Thus, entrainment affects the variability and large-scale energy trans-
547 port when coupled with AMOC.

548 CTL_EMOM produces SST variability patterns that are in better agreement with those of
 549 the OGCM in the mid-latitude North Pacific, South Pacific, and North Atlantic oceans. Improve-
 550 ments in surface air temperature variability over North America and the Southern Ocean are likely
 551 associated with improved SST variability located upstream. In the sea-ice loss experiment, the
 552 Ekman flow amplifies the ITCZ shift, which contradicts past literature. This disagreement is due
 553 to the detail of Ekman parameterization, i.e., the inclusion of $\epsilon\tau^y$ and the choice of $H_{\text{total}} = H_{\text{EK}} +$
 554 H_{RF} . Thus, the ITCZ response to sea-ice loss is sensitive to the Ekman parameterization.

555 In Section 2.6, we found a coupled positive feedback that originates from having insuffi-
 556 cient diffusion to counteract the convergence of the frictional-Ekman flow. The feedback excites
 557 a coupled mode in which the convergence in the wind field enhances the convergence of warm
 558 surface water, that feeds back positively on the atmospheric convection and associated low-level
 559 convergence. This feedback generates erroneous rainband (Figure S1a) on the equator in the Cen-
 560 tral Pacific where the cold tongues are still poorly simulated in many climate models (Tian & Dong,
 561 2020). It will be interesting in future studies to investigate whether this positive feedback and cold-
 562 tongue bias are connected.

563 Our hierarchy can still be refined depending on the need. For example:

- 564 • The nudging method used to derive the flux correction should account for the sea-ice con-
 565 centration and thickness in addition to SST. Using information about the sea-ice field will
 566 resolve the fluxes of heat that are not accompanied by a change of temperature because
 567 of the phase change. In addition, diagnosing the interannual variability of sea-ice form-
 568 ing heat flux $\overline{\Lambda'T'}$ in Equation (11) from OGCM and subtracting it out of the flux correc-
 569 tion may avoid double counting that causes the warm SST bias in the Southern Ocean along
 570 the sea-ice edges.
- 571 • The mixed-layer dynamics can further realize the K-profile parameterization (Large et al.,
 572 1994) or Niiler-Kraus type parameterization (Gaspar, 1988) so that it can respond dynam-
 573 ically to forcing. This improves the sub-seasonal to seasonal variability, which is impor-
 574 tant in the tropics.

Table 1. Sea-ice volume and area of CTL and SIL simulations in the northern and southern hemispheres.

	Sea-ice Volume [$\times 10^3 \text{ km}^3$]				Sea-ice Area [$\times 10^6 \text{ km}^2$]			
	NH		SH		NH		SH	
	CTL	SIL	CTL	SIL	CTL	SIL	CTL	SIL
OGCM	32.9	5.33	29.3	28.3	11.4	7.39	16.5	16.1
EMOM	26.2	5.20	21.0	19.5	10.8	7.02	14.4	13.7
MLM	26.5	5.21	23.6	23.1	10.8	7.02	15.1	14.9
SOM	28.7	5.25	30.4	30.1	11.3	7.15	16.9	16.8

- 575 • One can extend the current Ekman flow parameterization scheme by extending the thick-
- 576 ness of the return flow layer H_{RF} from a constant to a spatially varying variable to account
- 577 for sensitivity of OHT_{EK} to H_{RF} . Devising a framework to diagnose H_{RF} from an OGCM
- 578 will be necessary.
- 579 • Rayleigh friction is not the best parameterization to replace the momentum diffusion. One
- 580 can improve the Ekman flow parameterization using the original diffusion form $\partial_z(K_m \partial_z \vec{v})$.
- 581 Using this form leads to solving a fourth-order ordinary differential equation for each wa-
- 582 ter column.
- 583 • Ocean gyres are important for mid- to high-latitude OHT, but it is non-trivial to simulate
- 584 them directly. Still, if the gyre velocity does not change significantly, prescribing the ocean
- 585 flow can be a workaround to achieve responsive temperature and salt advection.

586 Appendix A Sea-ice forcing

587 We apply the forcing by constraining sea-ice thickness. The target sea-ice thickness in our

588 sea-ice thickness experiment is derived from years 2081–2100 of an ensemble member of Rep-

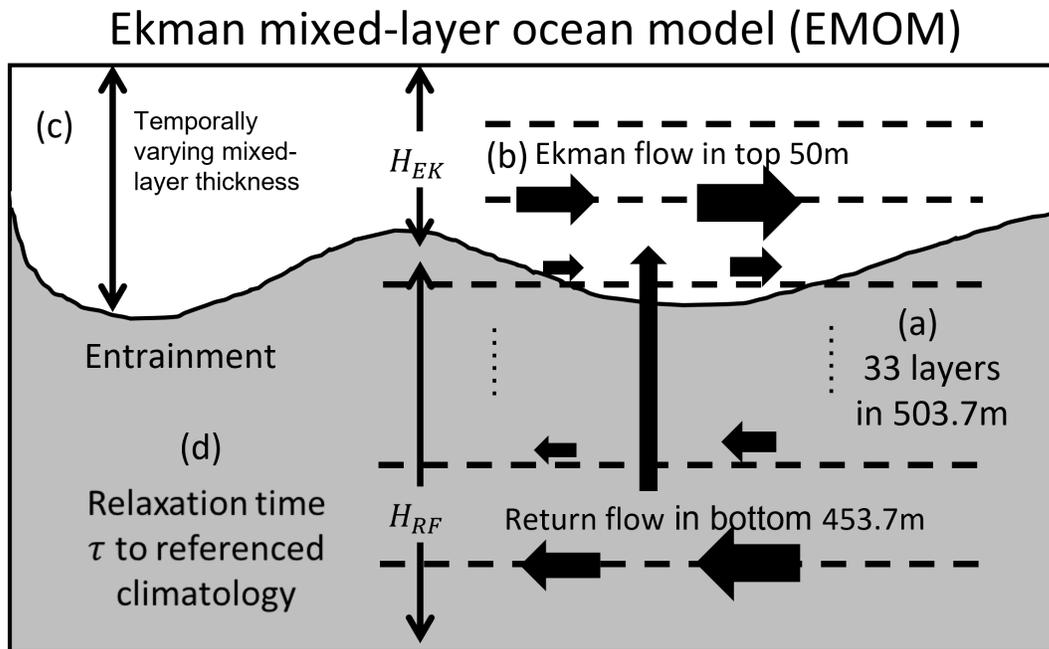


Figure 1. Schematic diagram of EMOM architecture. (a) EMOM has 33 vertical layers whose total depth is ≈ 503.7 m and range in thickness from 10 m to 48 m. (b) The Ekman and return flow layers have thicknesses of 50 m and 453.7 m. (c) EMOM has a time-varying mixed-layer thickness to capture the effect of seasonal entrainment and detrainment. (d) The ocean temperature and salinity are relaxed toward a reference three-dimensional profile with a 100-year timescale.

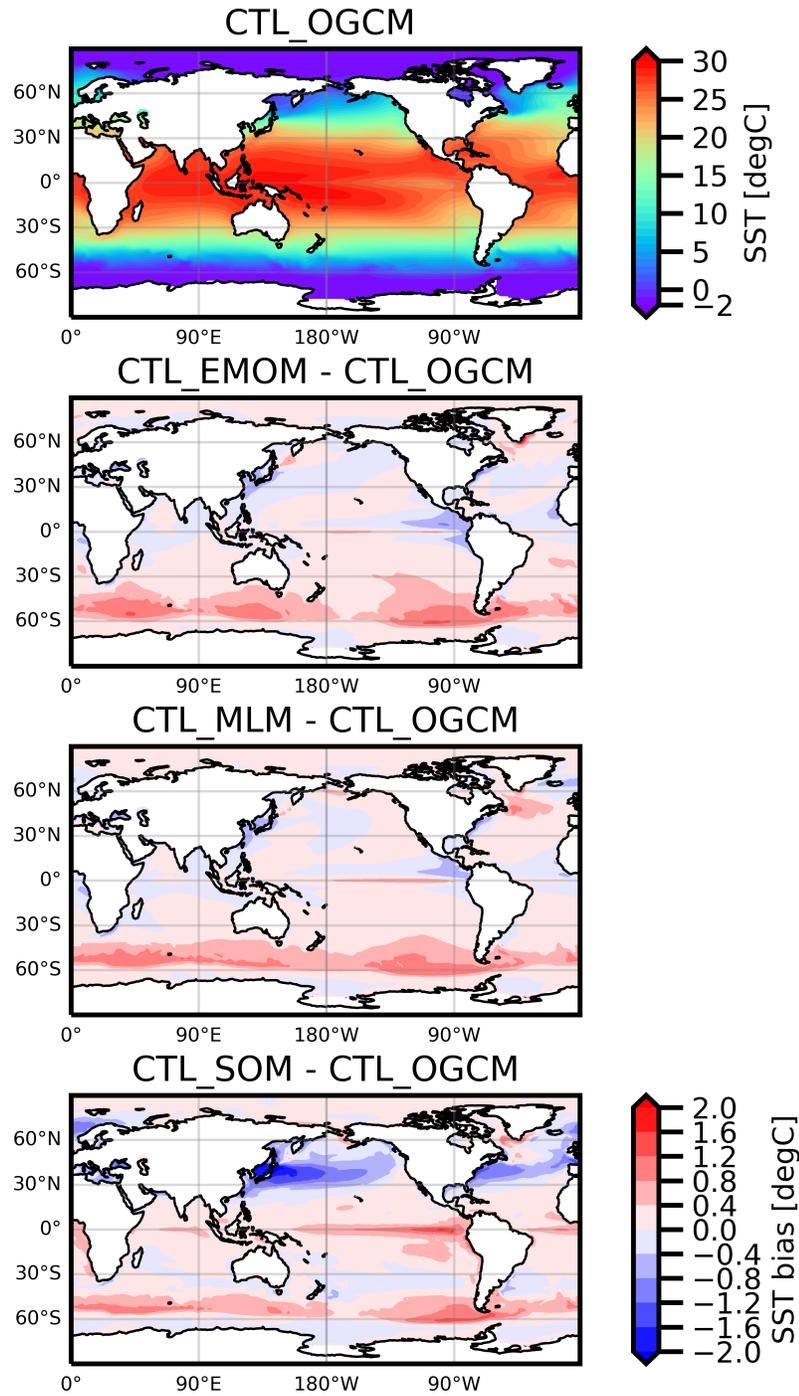


Figure 2. The annual mean SST of the hierarchy in CTL run. The top panel shows the target climatology obtained from the run using the full OGCM. The other panels show the biases in the other hierarchy members, i.e., $(\text{CTL}_{[\text{member}]} - \text{CTL}_{\text{OGCM}})$.

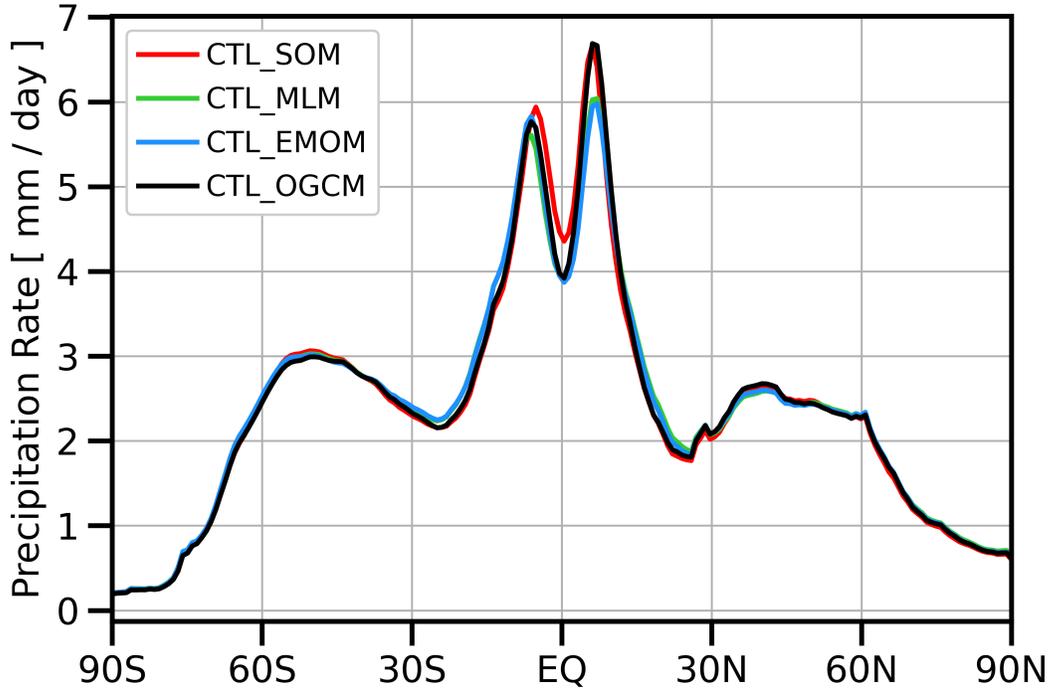


Figure 3. The zonal mean of the annually-averaged precipitation for the CTL run.

589 representative Concentration Pathway 8.5 (RCP8.5) in Coupled Model Intercomparison Project Phase
 590 5 (CMIP5) that is simulated using CCSM4. The total sea-ice volume in the SIL is about one-sixth
 591 of that in the CTL with an almost sea-ice-free Arctic in September. The distribution of mean sea-
 592 ice thickness is shown in Figure S3.

The target sea-ice thickness is achieved by imposing a pseudo heat flux to the sea-ice model,

$$F_{\text{pseudo}} = \frac{L_{\text{ice}}\rho_{\text{ice}}}{\tau_{\text{nudging}}} (h_{\text{model}} - h_{\text{target}}), \quad (\text{A1})$$

593 where h_{model} and h_{target} are the sea-ice thicknesses of the current time step and the target, F_{pseudo}
 594 is the pseudo heat flux with the sign chosen so that $F_{\text{pseudo}} > 0$ indicates an energy flux into the
 595 sea ice, $L_{\text{ice}} = 3.34 \times 10^5 \text{ J/kg}$ the specific latent heat fusion for sea ice, $\rho_{\text{ice}} = 917 \text{ kg/m}^3$ the
 596 sea ice density, $\tau_{\text{nudging}} = 5 \text{ days}$ is the nudging timescale, and h_{model} and h_{target} are the modeled
 597 and target sea-ice thicknesses. If the modeled sea-ice thickness is larger than the target sea-ice
 598 thickness then a heat gain causes the ice volume to decrease. Conversely, if the modeled sea-ice
 599 thickness is less than the target sea-ice thickness then a heat loss causes the ice volume to increase

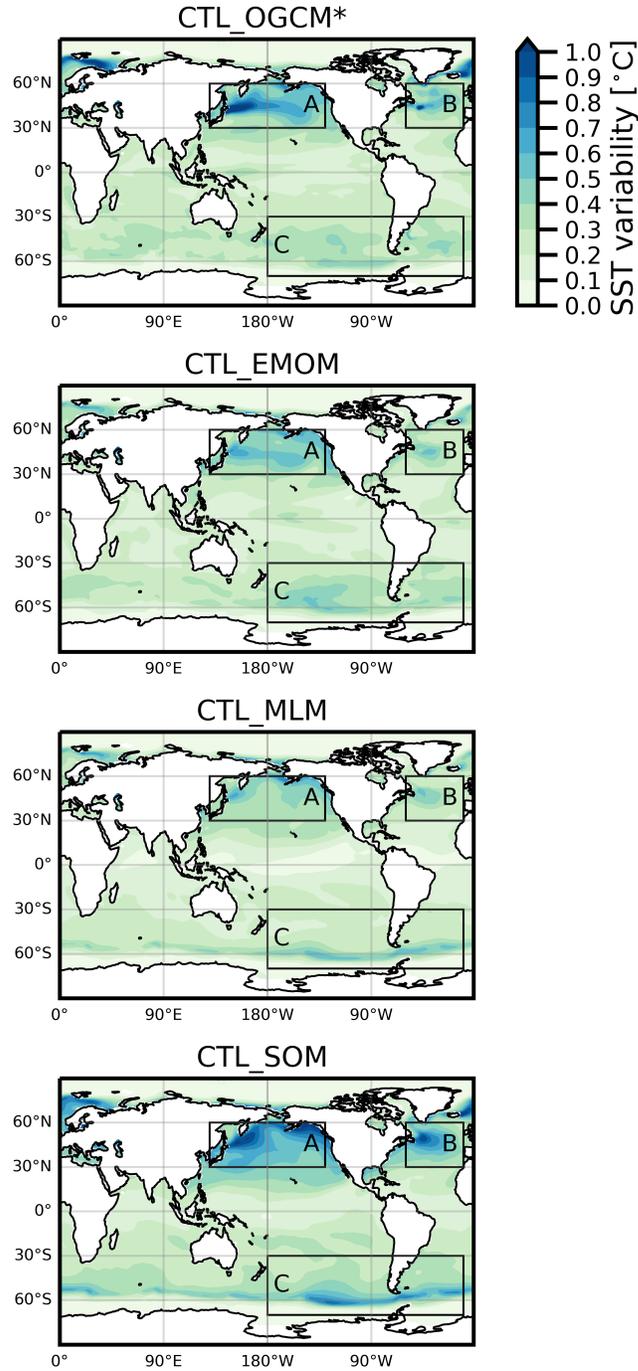


Figure 4. The annual mean SST variability (standard deviation of SSTA) of CTL run. The variability in CTL_OGCM* is computed by removing the anomalies that are correlated with the time series of the first empirical orthogonal function of monthly SSTA between 20°S–20°N. Boxes A, B, and C show the regions that have the biggest differences.

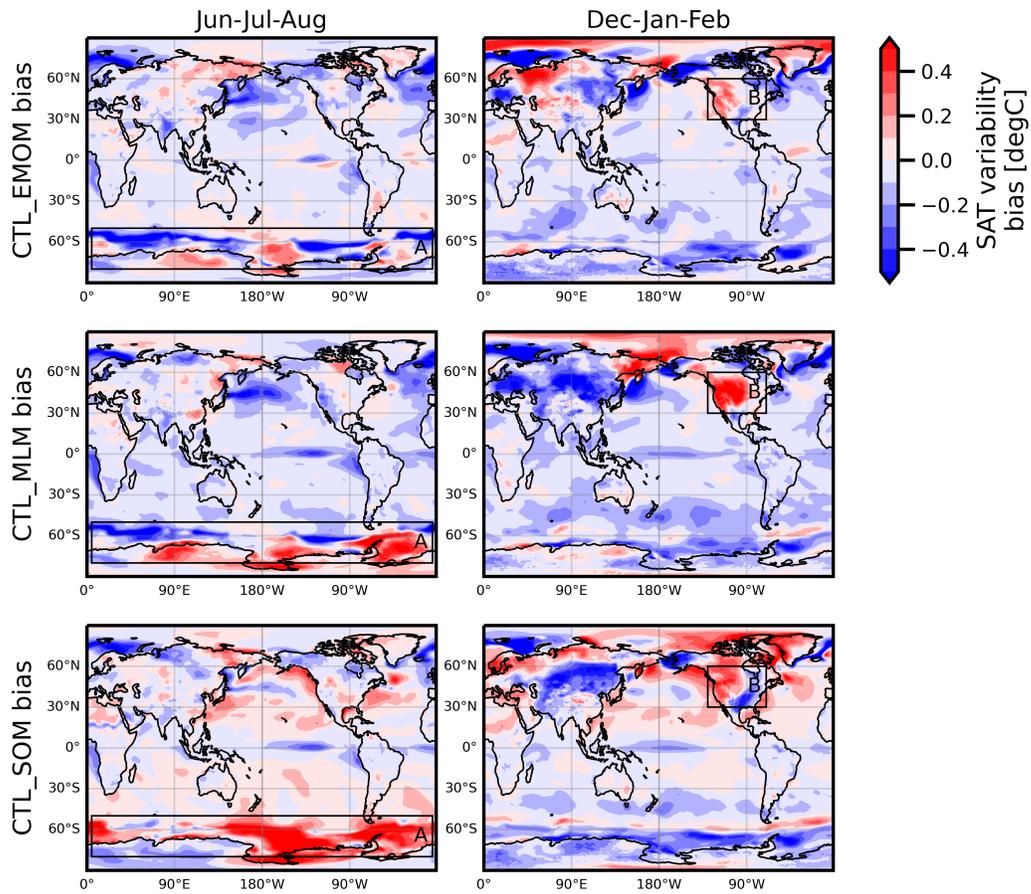


Figure 5. The Jun-Jul-Aug and Dec-Jan-Feb mean SAT variability bias (CTL_[member] - CTL_OGCM*).

See the caption for Figure 4 for the definition of CTL_OGCM*. Boxes A, B, and C show regions of major

SAT improvements that originate from the inclusion of Ekman flow.

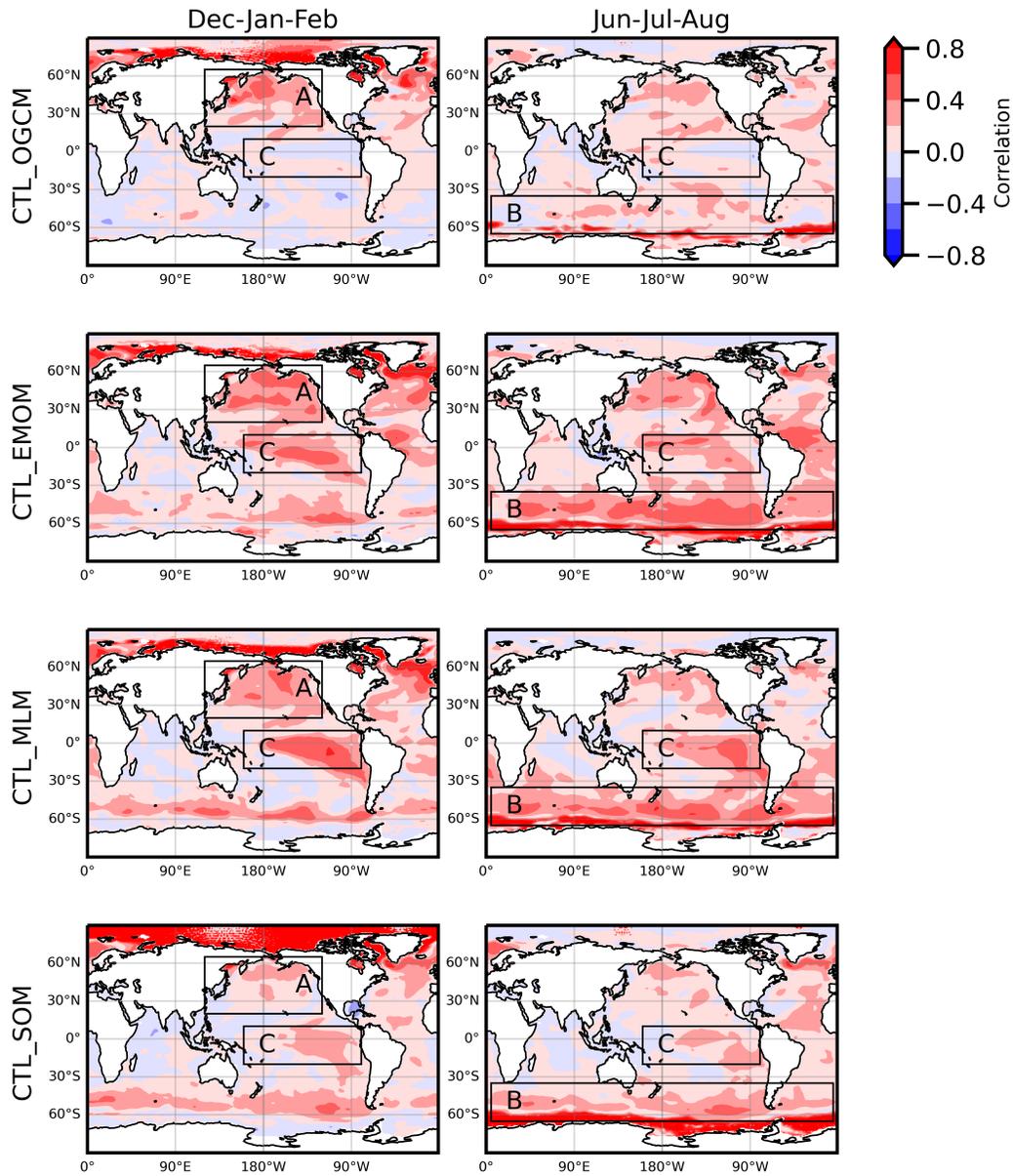


Figure 6. SSTA winter-to-winter correlation. Here Dec-Jan-Feb and Jun-Jul-Aug are winters in northern and southern hemispheres, respectively. Box A shows the improvement over the Northern Pacific as more ocean processes are included. Boxes B and C show that there are major processes not captured in simplified models that removes the memory. In box B, we speculate these are the northward transport of the subtended Antarctic Intermediate Water and the diffusive effect of eddies. In box C, horizontal currents and wave dynamics might act to dampen the SSTA efficiently.

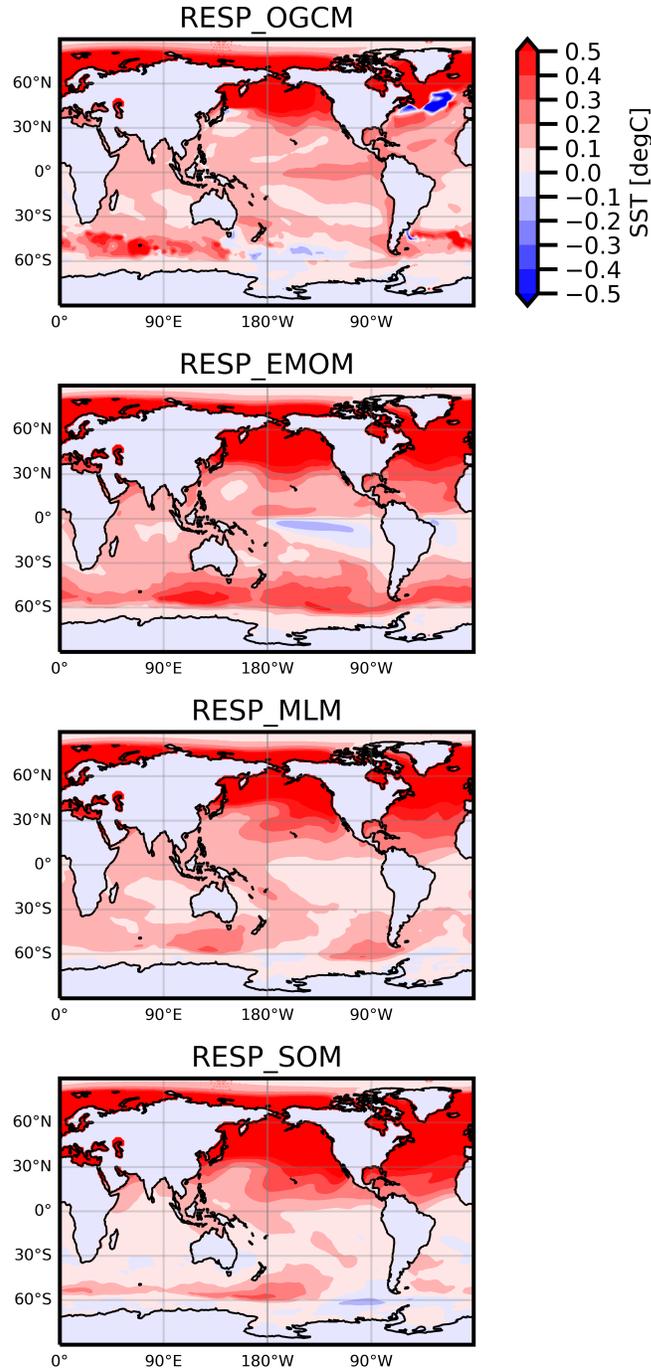


Figure 7. The annual mean SST response to Arctic sea-ice loss. The response is defined as $\text{RESP}_{[\text{member}]} := \text{SIL}_{[\text{member}]} - \text{CTL}_{[\text{member}]}$. As we move up the hierarchy, the extent of SST warming decreases, demonstrating the ocean’s improved ability to buffer the effect of the warming as more processes are included.

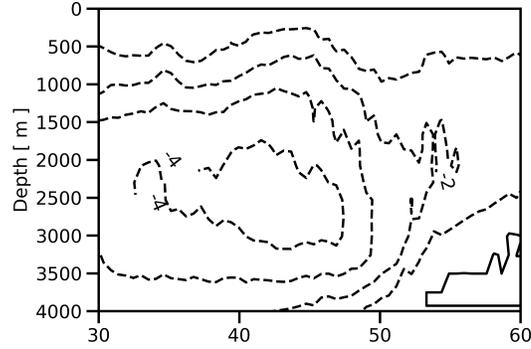


Figure 8. The AMOC streamfunction response to the loss of Arctic sea-ice. The contour spacing is $1 \text{ Sv} = 1 \times 10^6 \text{ m}^3/\text{s}$. The anomalous upwelling that removes the entrained heat from the surface is located between 40°N - 60°N . The heat removal creates the warming hole over the North Atlantic, as seen in Figure 7.

600 towards the target. The model we use divides the sea ice into five different categories with each
 601 having its own equivalent sea-ice thickness. Since we do not differentiate them, we use the value
 602 of their sum as sea-ice thickness in the equation above. Nudging the sea-ice thickness does not
 603 uniquely determine sea ice concentration which might give rise to inconsistent ice albedo and
 604 heat exchange if forcing is weak (Sun et al., 2020). Since our forcing is very strong the discrep-
 605 ancy is negligible.

606 **Appendix B Temporally Varying Mixed-layer Depth Changes SST variability**

607 We construct a one-dimensional mixed-layer model by vertically integrating Equation (1a)
 608 from the bottom of mixed-layer to the surface to get an approximated well-mixed mixed-layer model
 609 as

$$610 \quad \frac{\partial T_{\text{mix}}}{\partial t} = -\frac{F_{\text{net}}}{\rho c_p h} - \frac{w_e}{h} (T_{\text{mix}} - T_d) + q$$

611 where T_{mix} is the mixed-layer ocean temperature, $F_{\text{net}} = F(z=0) - F(z=-h)$ is the net incom-
 612 ing energy flux into the mixed-layer, T_d is the temperature immediately below the mixed-layer,
 613 $w_e = \max(\partial h/\partial t, 0)$ is the entrainment velocity, and q is the sum of freezing and advection ten-
 614 dency terms. Taking the deviation (denoted with a prime) from the seasonal mean (denoted with

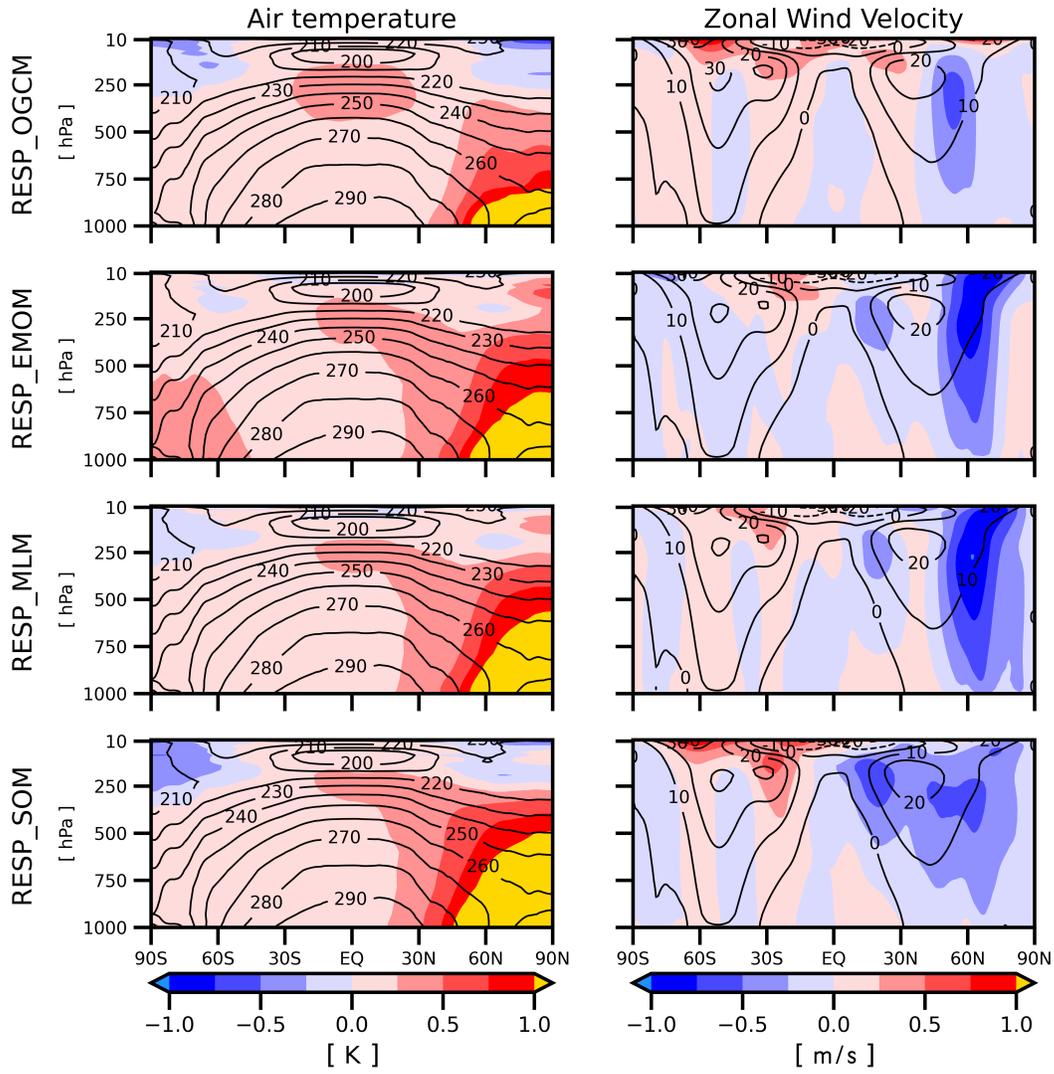


Figure 9. The zonal mean response of the atmosphere to Arctic sea-ice loss. The shading shows the response ($\text{RESP}_{[\text{member}]} := \text{SIL}_{[\text{member}]} - \text{CTL}_{[\text{member}]}$) of zonal mean air temperature (left column) and zonally mean zonal wind (right column). The contours show the quantities in $\text{CTL}_{[\text{member}]}$.

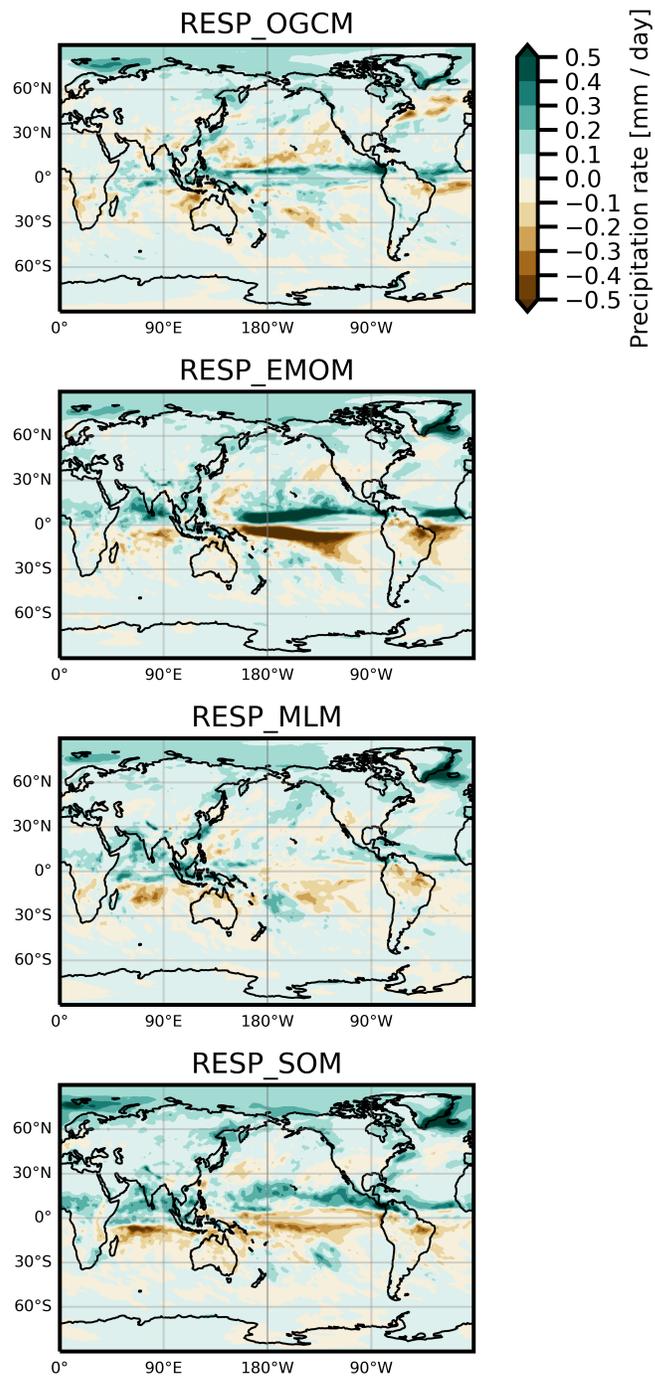


Figure 10. Similar to Figure 7 but for annual mean precipitation. Most of the responses locate in the tropics. RESP_OGCM shows a narrowing of ITCZ in the Pacific but a northward shift in the Atlantic. Rest of the models show various degree of northward shift of ITCZ.

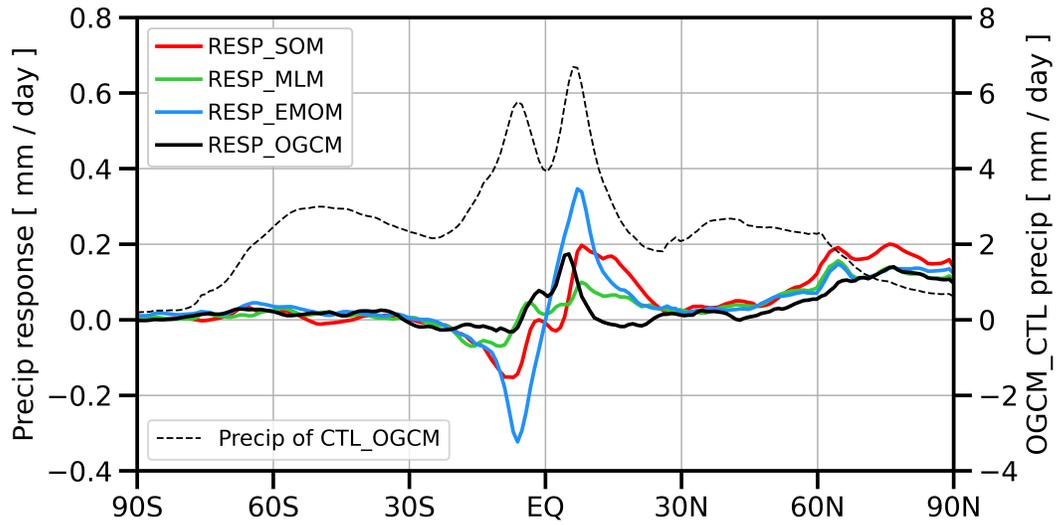


Figure 11. Response of zonal mean of annual precipitation ($\text{RESP}_{[\text{member}]} := \text{SIL}_{[\text{member}]} - \text{CTL}_{[\text{member}]}$; left ticks). Dashed line shows the precipitation in CTL_OGCM (right ticks).

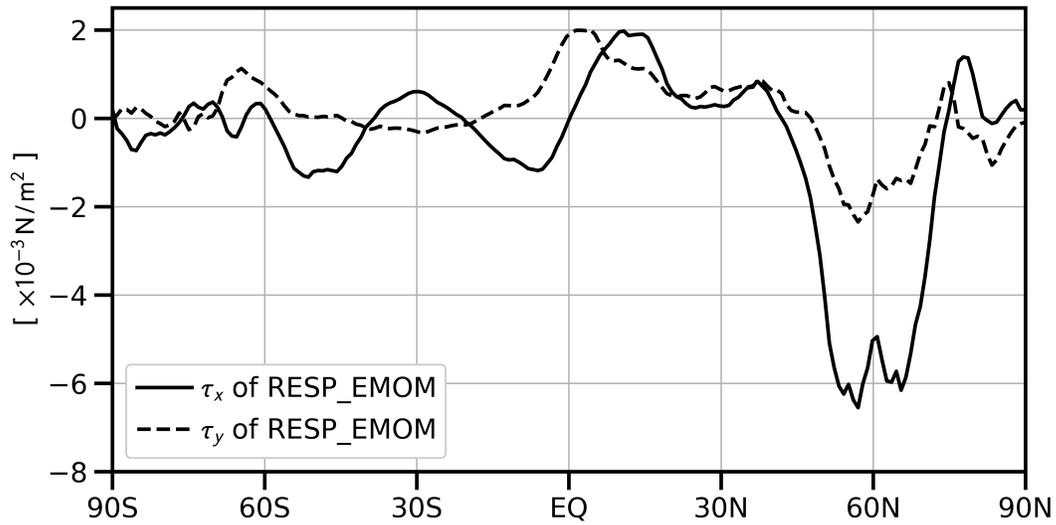


Figure 12. The response of surface zonal wind stress τ_x (solid; positive values means eastward wind anomalies) and meridional wind stress (dashed; positive values means northward wind anomalies) in EMOM.

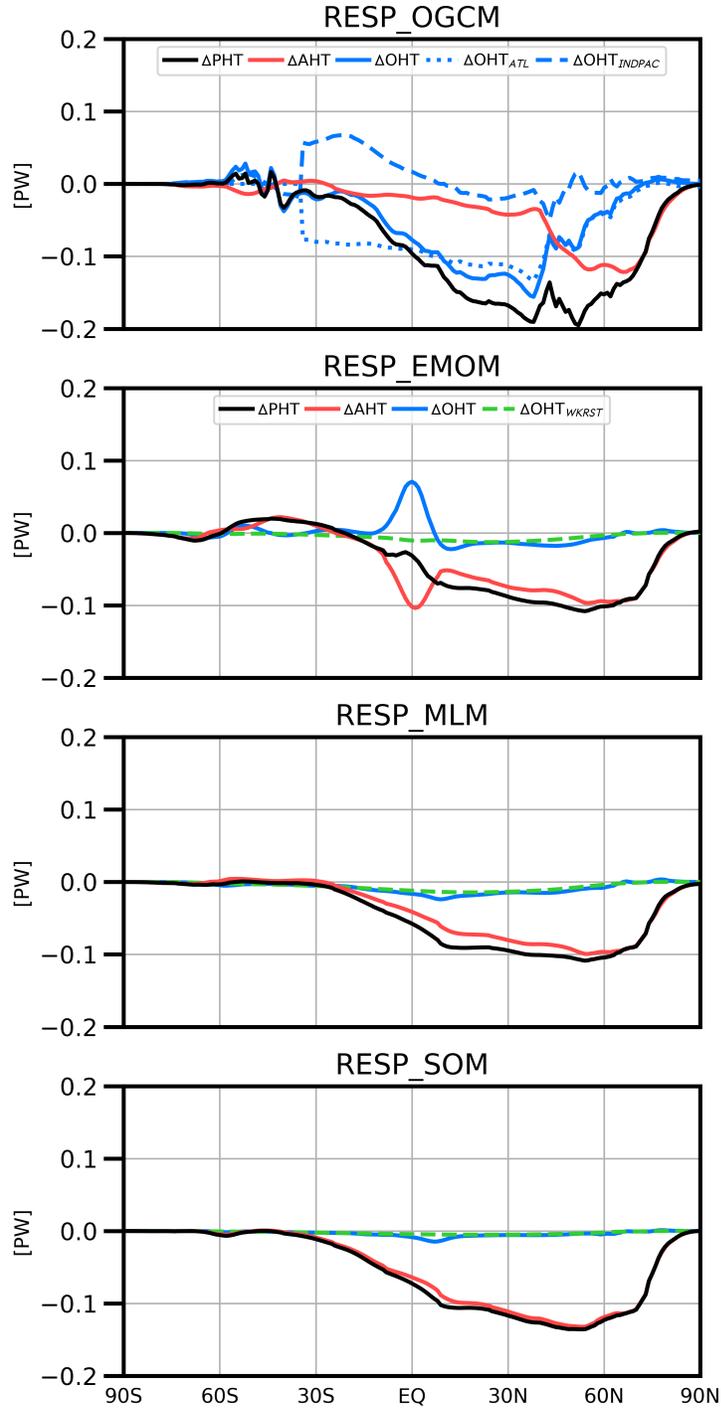


Figure 13. Heat transport response analysis. $PHT = AHT + OHT$. The OHT in RESP_OGCM is further decomposed into Indian-Pacific and Atlantic basins (ΔOHT_{INDPAC} and ΔOHT_{ATL}). In the simplified models, we also plot out the OHT due to weak-restoring (ΔOHT_{WKRST}) to show it is a minor component. In RESP_SOM and RESP_MLM, OHT is fully contributed by diffusion.

615 an overbar) we get

$$\begin{aligned}
 \frac{\partial T'_{\text{mix}}}{\partial t} &\approx -\frac{F'_{\text{net}}}{\rho c_p \bar{h}} + \frac{\bar{F}_{\text{net}}}{\rho c_p \bar{h}^2} h' - \left(\frac{w_e}{h}\right)' (\bar{T}_{\text{mix}} - \bar{T}_d) - \frac{\bar{w}_e}{h} (T'_{\text{mix}} - T'_d) + q' - \frac{1}{\tau_{\text{adj}}} T' \\
 &= -\frac{F'_{\text{net}}}{\rho c_p \bar{h}} - \frac{\bar{w}_e}{h} T' - \frac{1}{\tau_{\text{adj}}} T'
 \end{aligned} \tag{B1}$$

619 where the last term on the right-hand side is the damping of temperature anomaly due to air-sea
 620 interaction with time scale τ_{adj} . We drop the deviation terms related to h because we prescribed
 621 its annual evolution; we also drop q' because in the MLM model there is no advection and most
 622 of the ocean is sea-ice free; furthermore, we drop T_d for it is expected that $|T'_d| \ll |T'|$. We hy-
 623 pothesize that it is the existence of entrainment that reduces the SSTA variance.

624 Since strictly speaking, \bar{h} used in SOM is the temporal mean of \bar{h} in MLM so the first term
 625 on the right-hand side is also different, we should support this hypothesis by conducting stochas-
 626 tic simulations of Equation (B1).

627 We set

$$\bar{h}(t) = h_m + \frac{h_{\text{amp}}}{2} \sin\left(\frac{2\pi}{P}t\right) \tag{B2}$$

$$F'_{\text{net}}(t) = \left[F_m + \frac{F_{\text{amp}}}{2} \sin\left(\frac{2\pi}{P}t\right) \right] \epsilon'(t) \tag{B3}$$

631 where $P = 360$ days is the annual cycle period, $h_m = 70$ m is the mean mixed-layer thickness, h_{amp}
 632 is the amplitude of mixed-layer variation, $F_m = 30 \text{ W/m}^2$ and F_{amp} are the same but for F'_{net} , $\epsilon' \sim$
 633 N is a Gaussian noise. It is designed so such that the thicker mixed-layer thickness coincide with
 634 higher surface energy flux variance as observed in coupled simulations. SOM is represented by
 635 setting $h_{\text{amp}} = 0$ m and MLM by $h_{\text{amp}} = 60$ m. In all numerical integration we select $\tau_{\text{adj}} = 180$ days,
 636 $T'_{\text{mix}}(t=0) = 0$, and Euler forward scheme with time step 30 days is used. For each simulation,
 637 integrate for 100-thousand years and compute the resulting monthly mean values (30 days per
 638 month), then use the resulting T'_{mix} time series to compute one standard deviation value.

639 Assigning $F_{\text{amp}} = 0 \text{ W/m}^2$, the SST variability of SOM and MLM are 0.27°C and 0.21°C .
 640 Further assigning $F_{\text{amp}} = 40 \text{ W/m}^2$ we get 0.30°C and 0.20°C . The even larger separation is be-
 641 cause larger atmospheric stochastic forcing (winter) is efficiently damped by thicker mixed-layer

642 depth. These two sets experiments show that the main cause of lower SST variability in MLM
643 comes from the temporally varying mixed-layer depth.

644 **Open Research**

645 The simulation output data is available upon request. Please contact the authors through
646 the email address provided.

647 **Acknowledgments**

648 This work is supported by the Department of Energy Grant DE-SC0019407. Computational re-
649 sources were also provided by National Center for Atmospheric Research supercomputer Cheyenne.

650 **References**

- 651 Adam, O. (2021). Dynamic and energetic constraints on the modality and position of the in-
652 tertropical convergence zone in an aquaplanet. *Journal of Climate*, *34*(2), 527–543.
- 653 Afargan-Gerstman, H., & Adam, O. (2020). Nonlinear damping of itcz migrations due to ek-
654 man ocean energy transport. *Geophysical Research Letters*, *47*(5), e2019GL086445.
- 655 Alexander, M. A., & Deser, C. (1995). A mechanism for the recurrence of wintertime mid-
656 latitude sst anomalies. *Journal of physical oceanography*, *25*(1), 122–137.
- 657 Alexander, M. A., Deser, C., & Timlin, M. S. (1999). The reemergence of sst anomalies in
658 the north pacific ocean. *Journal of climate*, *12*(8), 2419–2433.
- 659 Alexander, M. A., & Scott, J. D. (2008). The role of ekman ocean heat transport in the north-
660 ern hemisphere response to enso. *Journal of climate*, *21*(21), 5688–5707.
- 661 Alexander, M. A., Scott, J. D., & Deser, C. (2000). Processes that influence sea surface tem-
662 perature and ocean mixed layer depth variability in a coupled model. *Journal of Geo-
663 physical Research: Oceans*, *105*(C7), 16823–16842.
- 664 Bitz, C. M., Shell, K., Gent, P., Bailey, D., Danabasoglu, G., Armour, K., . . . Kiehl, J.
665 (2012). Climate sensitivity of the community climate system model, version 4. *Jour-
666 nal of Climate*, *25*(9), 3053–3070.

- 667 Bryan, K., & Cox, M. D. (1972). An approximate equation of state for numerical models of
 668 ocean circulation. *J. Phys. Oceanogr*, 2(4), 510–514.
- 669 Claussen, M., Mysak, L., Weaver, A., Crucifix, M., Fichet, T., Loutre, M.-F., . . . others
 670 (2002). Earth system models of intermediate complexity: closing the gap in the
 671 spectrum of climate system models. *Climate dynamics*, 18(7), 579–586.
- 672 Codron, F. (2012). Ekman heat transport for slab oceans. *Climate Dynamics*, 38(1-2), 379–
 673 389.
- 674 Danabasoglu, G., Bates, S. C., Briegleb, B. P., Jayne, S. R., Jochum, M., Large, W. G., . . .
 675 Yeager, S. G. (2012). The cesm4 ocean component. *Journal of Climate*, 25(5),
 676 1361–1389.
- 677 Donohoe, A., Frierson, D. M., & Battisti, D. S. (2014). The effect of ocean mixed layer depth
 678 on climate in slab ocean aquaplanet experiments. *Climate dynamics*, 43(3-4), 1041–
 679 1055.
- 680 Gaspar, P. (1988). Modeling the seasonal cycle of the upper ocean. *Journal of Physical*
 681 *Oceanography*, 18(2), 161–180.
- 682 Gent, P. R., Danabasoglu, G., Donner, L. J., Holland, M. M., Hunke, E. C., Jayne, S. R., . . .
 683 others (2011). The community climate system model version 4. *Journal of climate*,
 684 24(19), 4973–4991.
- 685 Gent, P. R., & McWilliams, J. C. (1990). Isopycnal mixing in ocean circulation models.
 686 *Journal of Physical Oceanography*, 20(1), 150–155.
- 687 Green, B., & Marshall, J. (2017). Coupling of trade winds with ocean circulation damps itcz
 688 shifts. *Journal of Climate*, 30(12), 4395–4411.
- 689 Green, B., Marshall, J., & Campin, J.-M. (2019). The ‘sticky’ itcz: ocean-moderated itcz
 690 shifts. *Climate dynamics*, 53(1-2), 1–19.
- 691 Haarsma, R. J., Campos, E. J., Hazeleger, W., Severijns, C., Piola, A. R., & Molteni, F.
 692 (2005). Dominant modes of variability in the south atlantic: A study with a hierarchy
 693 of ocean–atmosphere models. *Journal of climate*, 18(11), 1719–1735.
- 694 Held, I. M. (2001). The partitioning of the poleward energy transport between the tropical

- 695 ocean and atmosphere. *Journal of the atmospheric sciences*, 58(8), 943–948.
- 696 Held, I. M. (2005). The gap between simulation and understanding in climate modeling. *Bul-*
 697 *letin of the American Meteorological Society*, 86(11), 1609–1614.
- 698 Hirons, L., Klingaman, N., & Woolnough, S. (2015). Metum-goml: a near-globally coupled
 699 atmosphere–ocean-mixed-layer model. *Geoscientific Model Development*, 8, 363–379.
- 700 Hunke, E. C., Lipscomb, W. H., Turner, A. K., Jeffery, N., & Elliott, S. (2010). Cice:
 701 the los alamos sea ice model documentation and software user’s manual version 4.1
 702 la-cc-06-012. *T-3 Fluid Dynamics Group, Los Alamos National Laboratory*, 675, 500.
- 703 Hurrell, J. W., Holland, M. M., Gent, P. R., Ghan, S., Kay, J. E., Kushner, P. J., . . . others
 704 (2013). The community earth system model: a framework for collaborative research.
 705 *Bulletin of the American Meteorological Society*, 94(9), 1339–1360.
- 706 Jayne, S. R., & Marotzke, J. (2001). The dynamics of ocean heat transport variability. *Re-*
 707 *views of Geophysics*, 39(3), 385–411.
- 708 Jeevanjee, N., Hassanzadeh, P., Hill, S., & Sheshadri, A. (2017). A perspective on climate
 709 model hierarchies. *Journal of Advances in Modeling Earth Systems*, 9(4), 1760–1771.
- 710 Kang, S. M. (2020). Extratropical influence on the tropical rainfall distribution. *Current Cli-*
 711 *mate Change Reports*, 6(1), 24–36.
- 712 Kang, S. M., Shin, Y., & Codron, F. (2018). The partitioning of poleward energy transport
 713 response between the atmosphere and ekman flux to prescribed surface forcing in a
 714 simplified gcm. *Geoscience Letters*, 5(1), 22.
- 715 Kang, S. M., Shin, Y., & Xie, S.-P. (2018). Extratropical forcing and tropical rainfall dis-
 716 tribution: energetics framework and ocean ekman advection. *npj Climate and Atmo-*
 717 *spheric Science*, 1(1), 1–10.
- 718 Kiehl, J. T., Shields, C. A., Hack, J. J., & Collins, W. D. (2006). The climate sensitivity of
 719 the community climate system model version 3 (ccsm3). *Journal of Climate*, 19(11),
 720 2584–2596.
- 721 Kwon, Y.-O., Deser, C., & Cassou, C. (2011). Coupled atmosphere–mixed layer ocean re-
 722 sponse to ocean heat flux convergence along the kuroshio current extension. *Climate*

- 723 *dynamics*, 36(11-12), 2295–2312.
- 724 Large, W. G., McWilliams, J. C., & Doney, S. C. (1994). Oceanic vertical mixing: A review
725 and a model with a nonlocal boundary layer parameterization. *Reviews of Geophysics*,
726 32(4), 363–403.
- 727 Lawrence, D. M., Oleson, K. W., Flanner, M. G., Thornton, P. E., Swenson, S. C., Lawrence,
728 P. J., . . . others (2011). Parameterization improvements and functional and structural
729 advances in version 4 of the community land model. *Journal of Advances in Modeling
730 Earth Systems*, 3(1).
- 731 Lee, T., & Marotzke, J. (1998). Seasonal cycles of meridional overturning and heat transport
732 of the indian ocean. *Journal of Physical Oceanography*, 28(5), 923–943.
- 733 Leonard, B. P. (1979). A stable and accurate convective modelling procedure based on
734 quadratic upstream interpolation. *Computer methods in applied mechanics and engi-
735 neering*, 19(1), 59–98.
- 736 Maher, P., Gerber, E. P., Medeiros, B., Merlis, T. M., Sherwood, S., Sheshadri, A., . . .
737 Zurita-Gotor, P. (2019). Model hierarchies for understanding atmospheric circula-
738 tion. *Reviews of Geophysics*, 57(2), 250–280.
- 739 Neale, R. B., Richter, J., Park, S., Lauritzen, P. H., Vavrus, S. J., Rasch, P. J., & Zhang, M.
740 (2013). The mean climate of the community atmosphere model (cam4) in forced sst
741 and fully coupled experiments. *Journal of Climate*, 26(14), 5150–5168.
- 742 Newman, M., Alexander, M. A., Ault, T. R., Cobb, K. M., Deser, C., Di Lorenzo, E., . . .
743 others (2016). The pacific decadal oscillation, revisited. *Journal of Climate*, 29(12),
744 4399–4427.
- 745 Niiler, P. (1977). One-dimensional models of the upper ocean. modelling and prediction of
746 the upper layers of the ocean, eb kraus, ed. *Pergamon Press*, 143, 172.
- 747 Nummelin, A., Busecke, J. J., Haine, T. W., & Abernathey, R. P. (2021). Diagnosing the
748 scale-and space-dependent horizontal eddy diffusivity at the global surface ocean.
749 *Journal of Physical Oceanography*, 51(2), 279–297.
- 750 Paulson, C. A., & Simpson, J. J. (1977). Irradiance measurements in the upper ocean. *Jour-*

- 751 *nal of Physical Oceanography*, 7(6), 952–956.
- 752 Schneider, T. (2017). Feedback of atmosphere-ocean coupling on shifts of the intertropical
753 convergence zone. *Geophysical Research Letters*, 44(22), 11–644.
- 754 Schneider, T., Bischoff, T., & Haug, G. H. (2014). Migrations and dynamics of the intertrop-
755 ical convergence zone. *Nature*, 513(7516), 45–53.
- 756 Screen, J. A., Deser, C., Smith, D. M., Zhang, X., Blackport, R., Kushner, P. J., . . . Sun, L.
757 (2018). Consistency and discrepancy in the atmospheric response to arctic sea-ice loss
758 across climate models. *Nature Geoscience*, 11(3), 155.
- 759 Smith, D. M., Eade, R., M., A., H., A., A., C., S., C., . . . others, W. A. (2022). Robust
760 but weak winter atmospheric circulation response to future arctic sea ice loss. , ac-
761 ceptedthe double-itcz bias in cmip3, cmip5, and cmip6 models based on annual mean
762 precipitation. *Nature Communications*.
- 763 Smith, R., Jones, P., Briegleb, B., Bryan, F., Danabasoglu, G., Dennis, J., . . . others (2010).
764 The parallel ocean program (pop) reference manual: ocean component of the com-
765 munity climate system model (ccsm) and community earth system model (cesm).
766 *LAUR-01853*, 141, 1–140.
- 767 Sun, L., Deser, C., Tomas, R. A., & Alexander, M. (2020). Global coupled climate re-
768 sponse to polar sea ice loss: Evaluating the effectiveness of different ice-constraining
769 approaches. *Geophysical Research Letters*, 47(3), e2019GL085788.
- 770 Tian, B., & Dong, X. (2020). The double-itcz bias in cmip3, cmip5, and cmip6 models based
771 on annual mean precipitation. *Geophysical Research Letters*, 47(8), e2020GL087232.
- 772 Tomas, R. A., Deser, C., & Sun, L. (2016). The role of ocean heat transport in the global cli-
773 mate response to projected arctic sea ice loss. *Journal of Climate*, 29(19), 6841–6859.
- 774 Vallis, G. K., Colyer, G., Geen, R., Gerber, E., Jucker, M., Maher, P., . . . Thomson, S. I.
775 (2018). Isca, v1. 0: A framework for the global modelling of the atmospheres of earth
776 and other planets at varying levels of complexity. *Geoscientific Model Development*,
777 11(3), 843–859.
- 778 Wang, Z., Schneider, E. K., & Burls, N. J. (2019). The sensitivity of climatological sst to

slab ocean model thickness. *Climate Dynamics*, 53(9), 5709–5723.

Figure 1.

Ekman mixed-layer ocean model (EMOM)

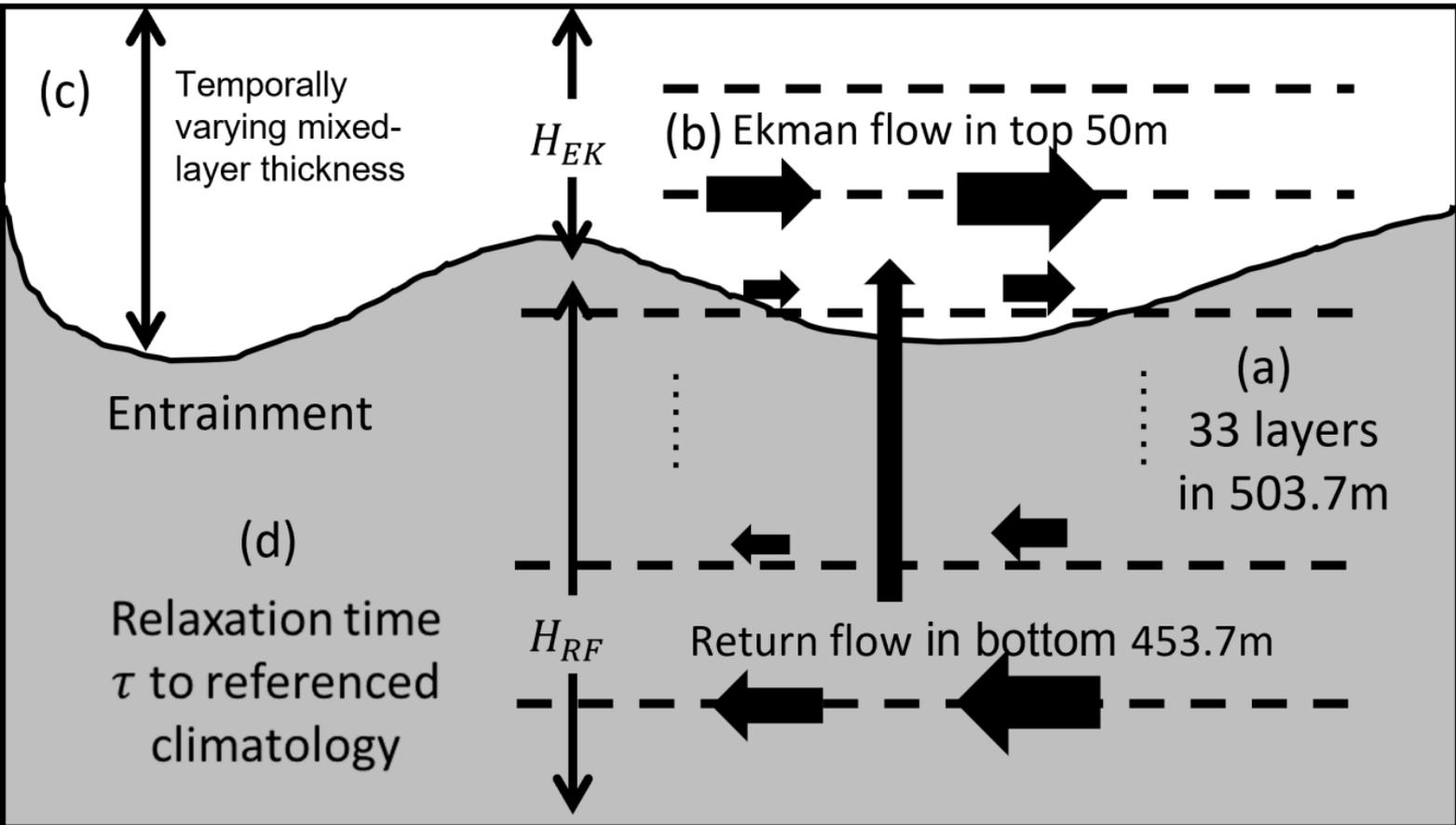
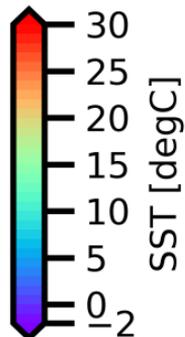
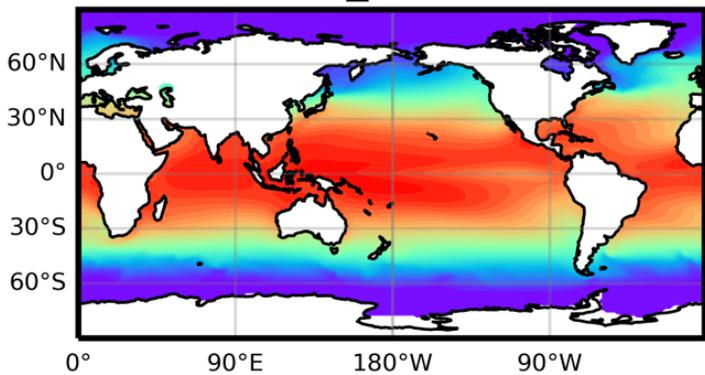
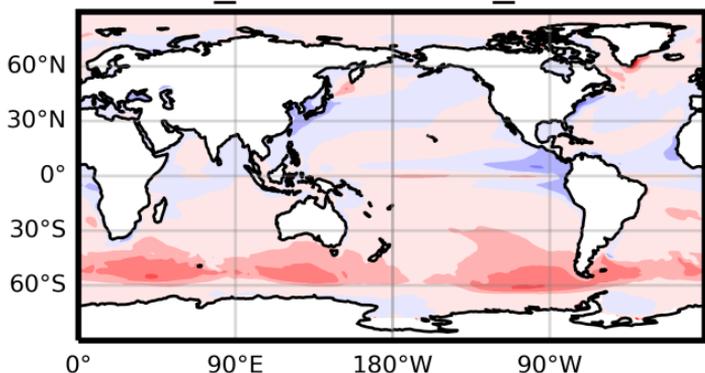


Figure 2.

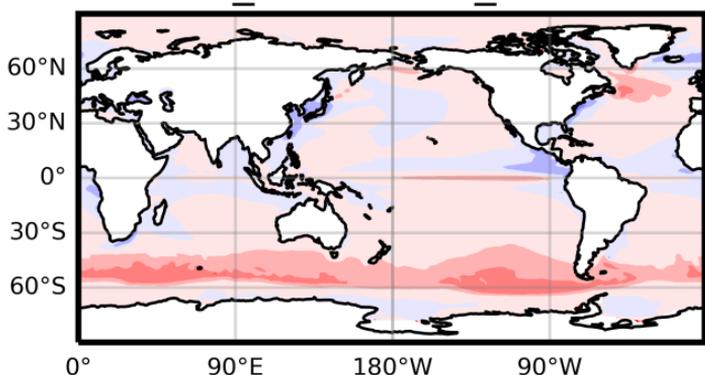
CTL_OGCM



CTL_EMOM - CTL_OGCM



CTL_MLM - CTL_OGCM



CTL_SOM - CTL_OGCM

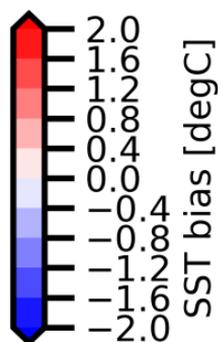
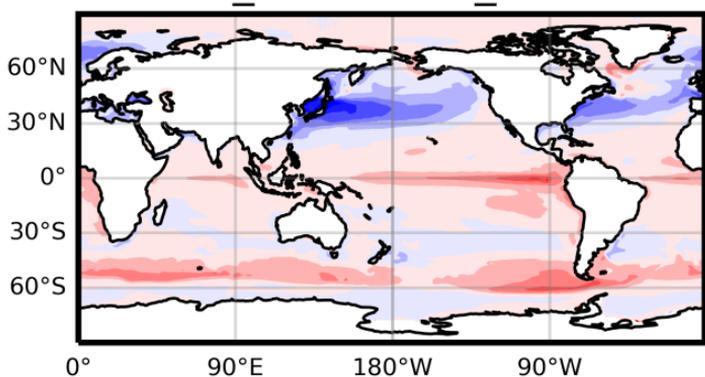


Figure 3.

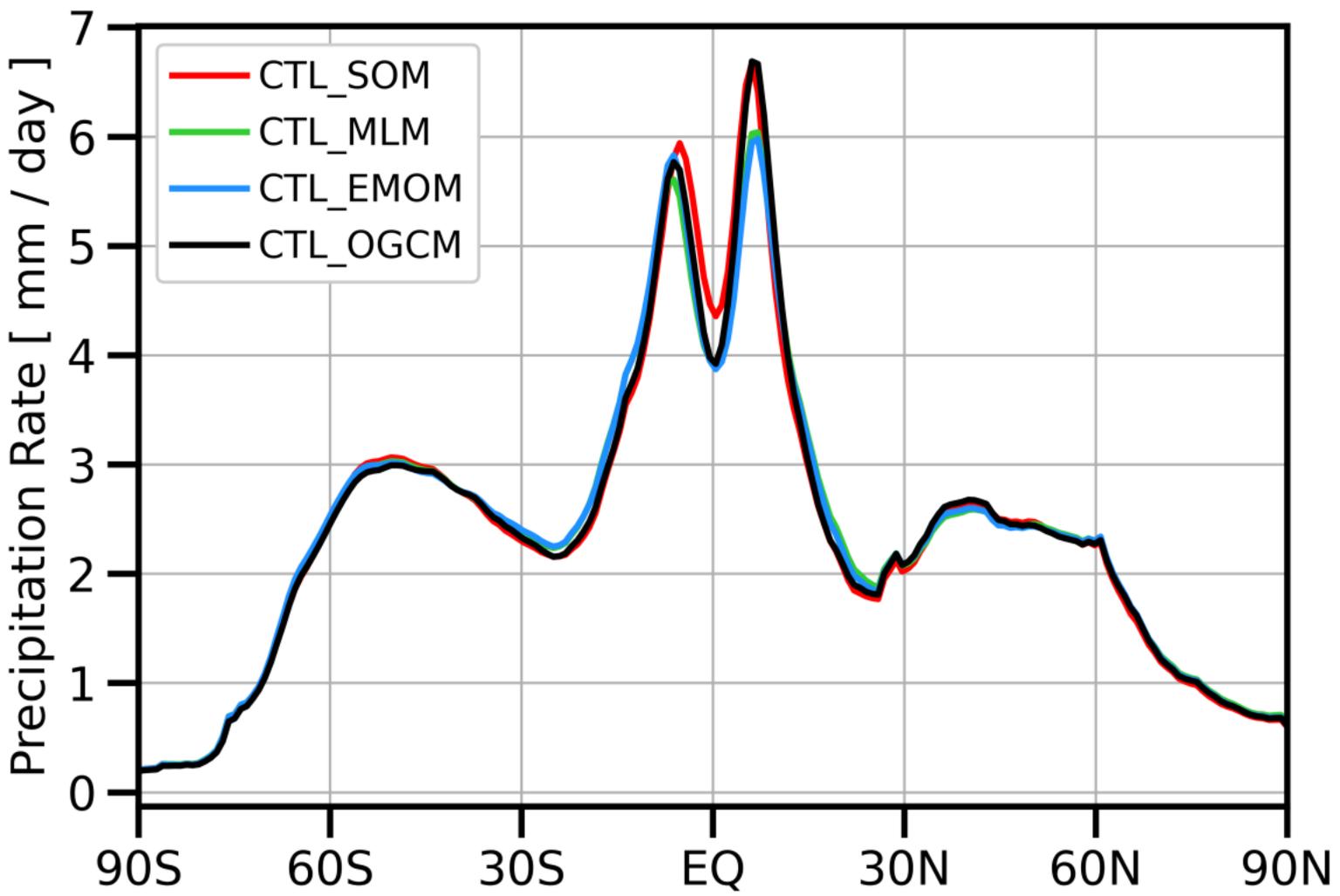
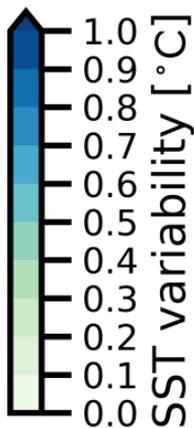
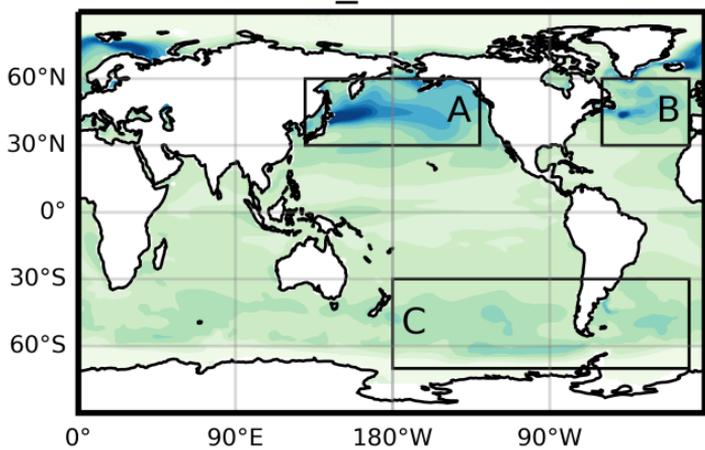
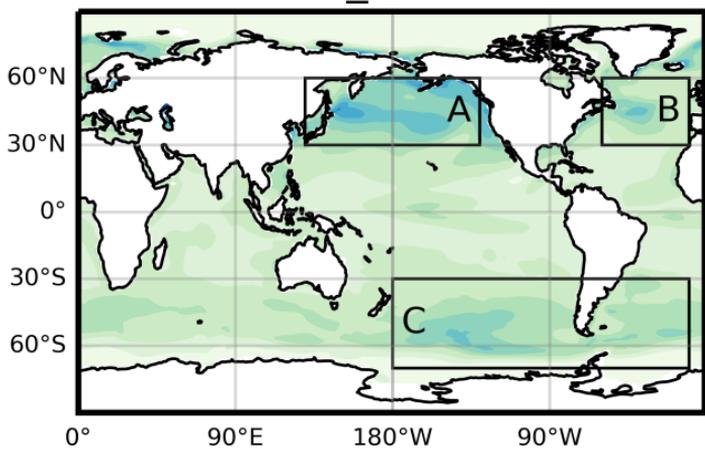


Figure 4.

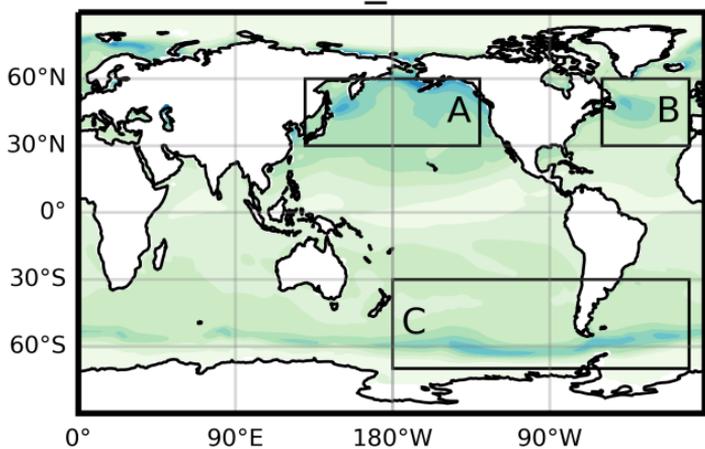
CTL_OGCM*



CTL_EMOM



CTL_MLM



CTL_SOM

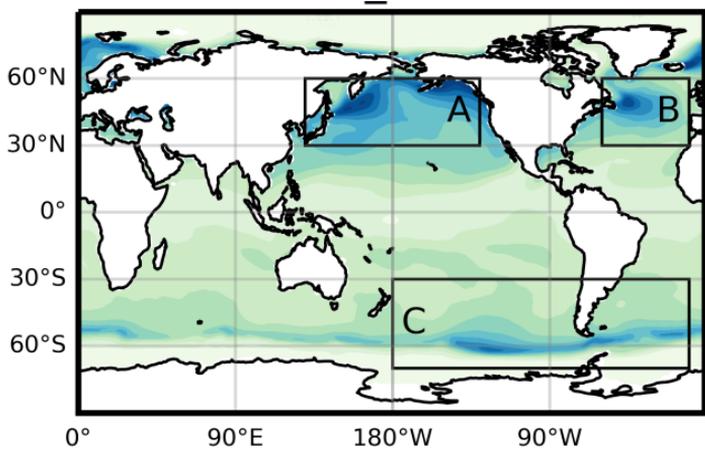
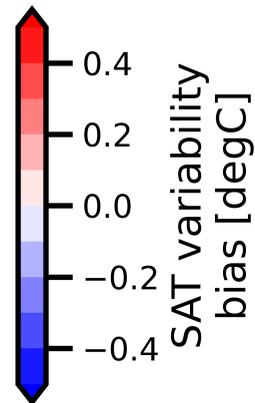
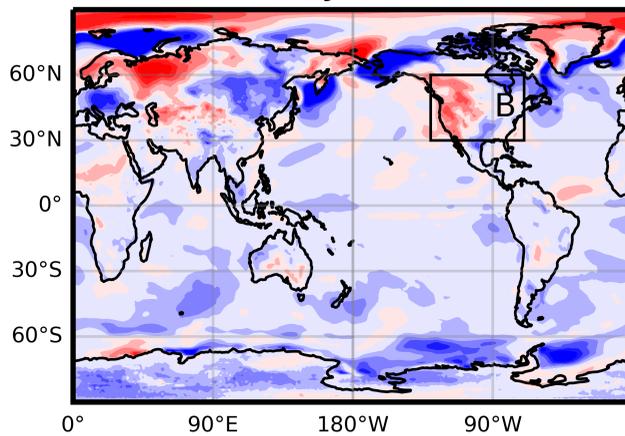
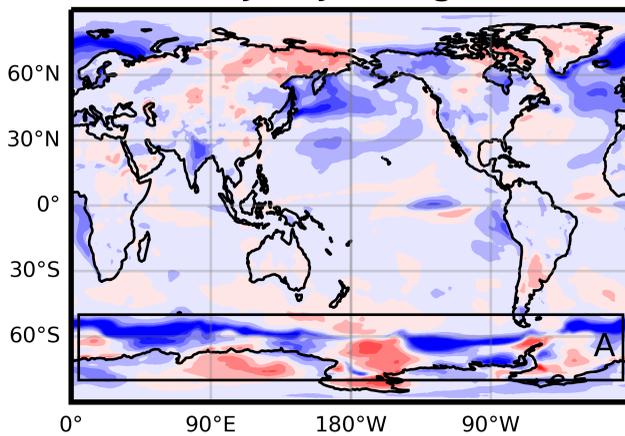


Figure 5.

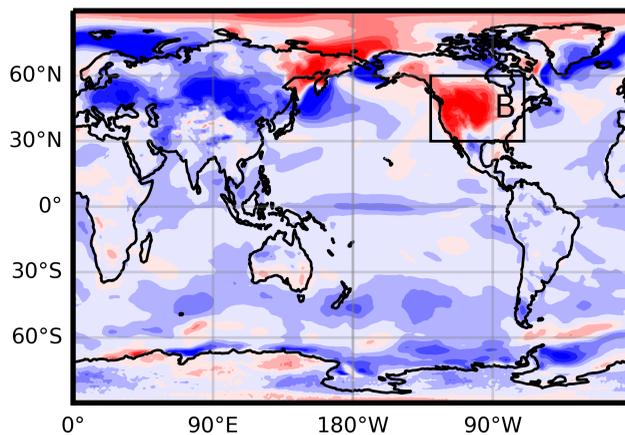
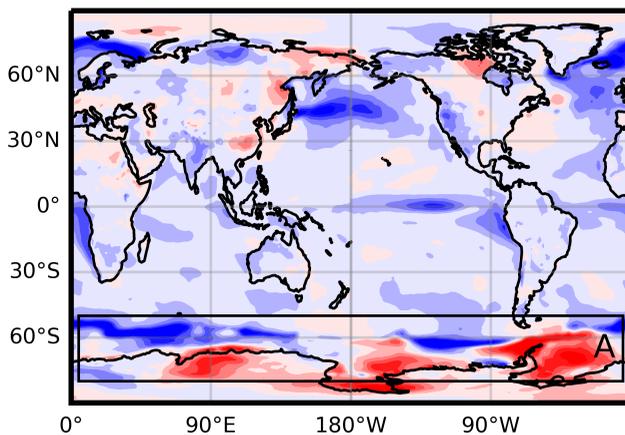
Jun-Jul-Aug

Dec-Jan-Feb

CTL_EMOM bias



CTL_MLM bias



CTL_SOM bias

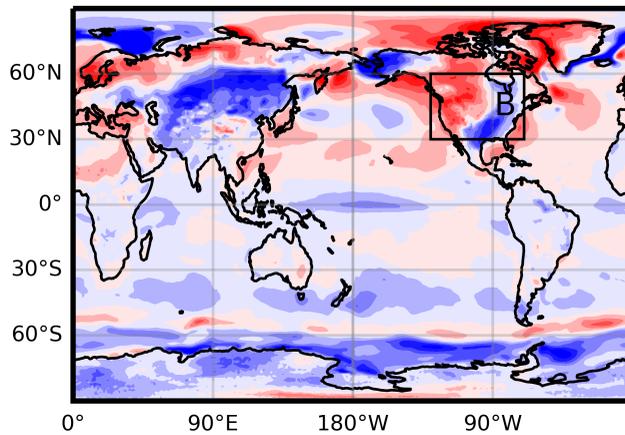
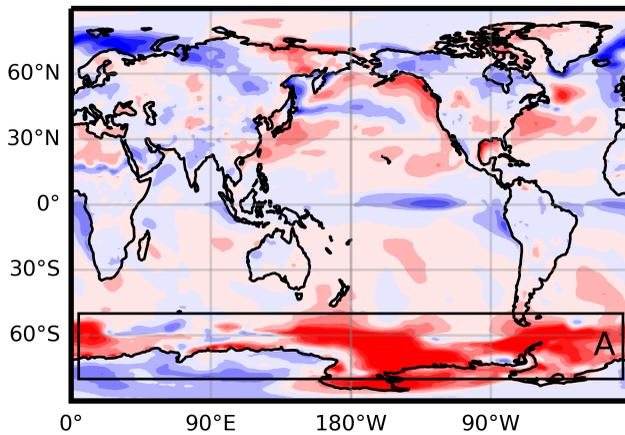
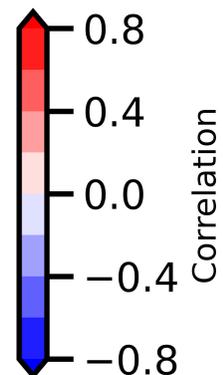


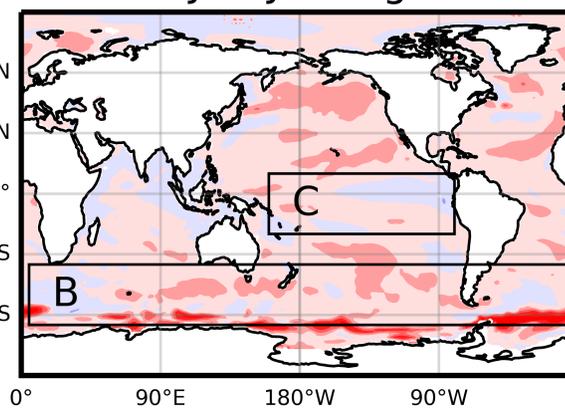
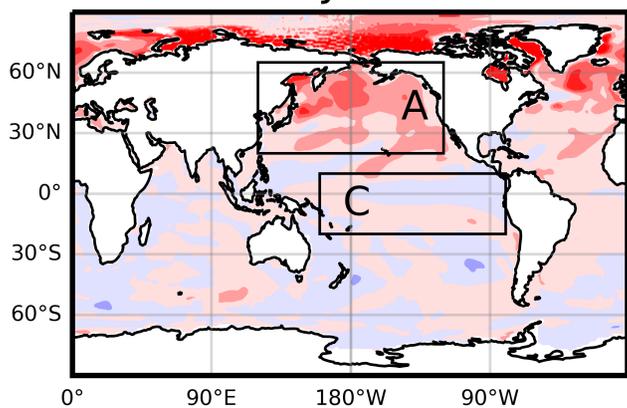
Figure 6.

Dec-Jan-Feb

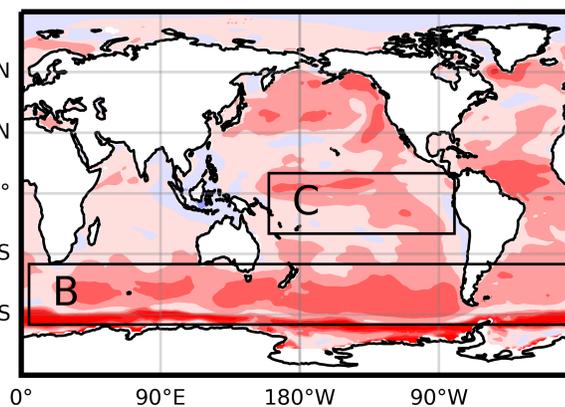
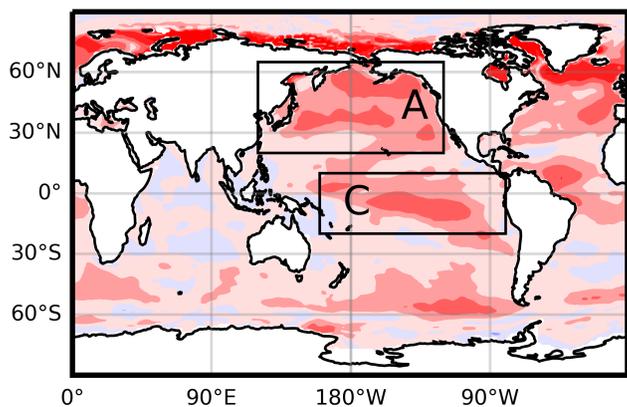
Jun-Jul-Aug



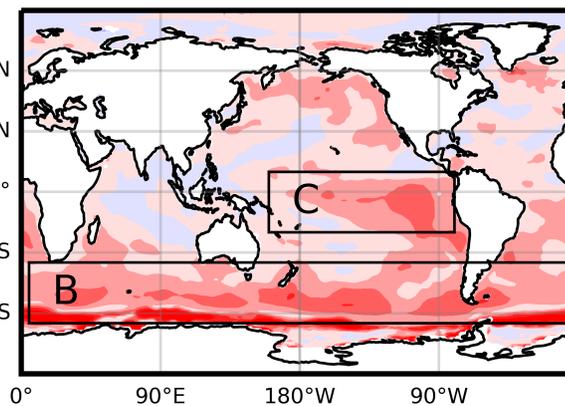
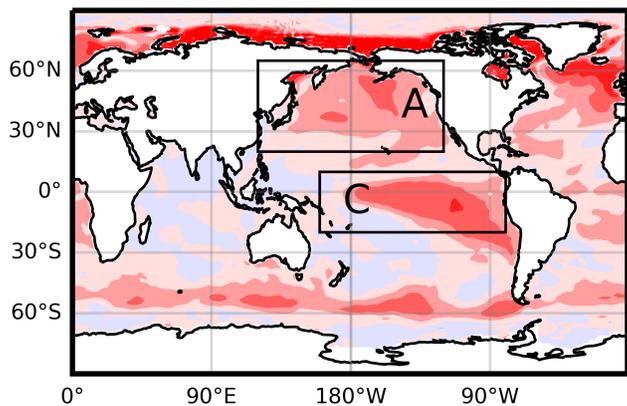
CTL_OGCM



CTL_EMOM



CTL_MLM



CTL_SOM

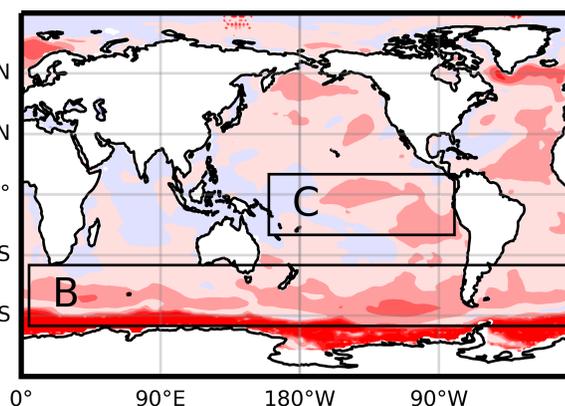
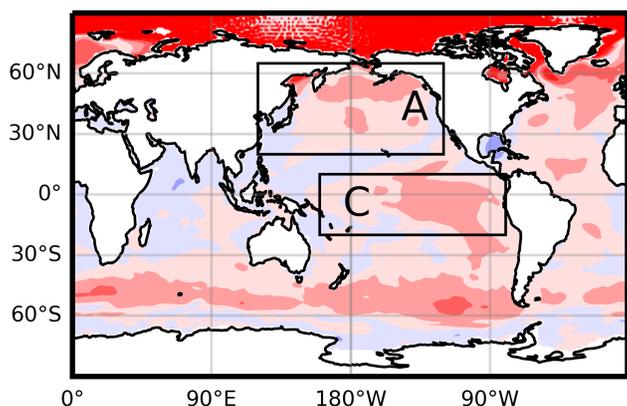
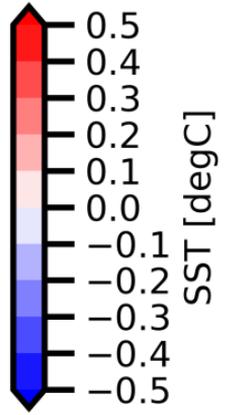
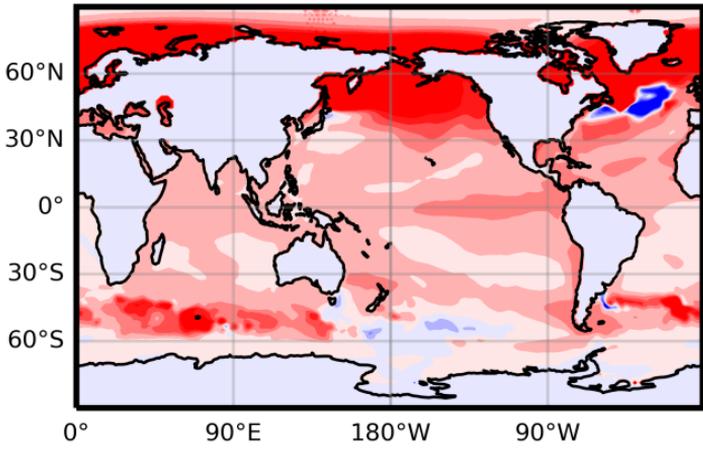
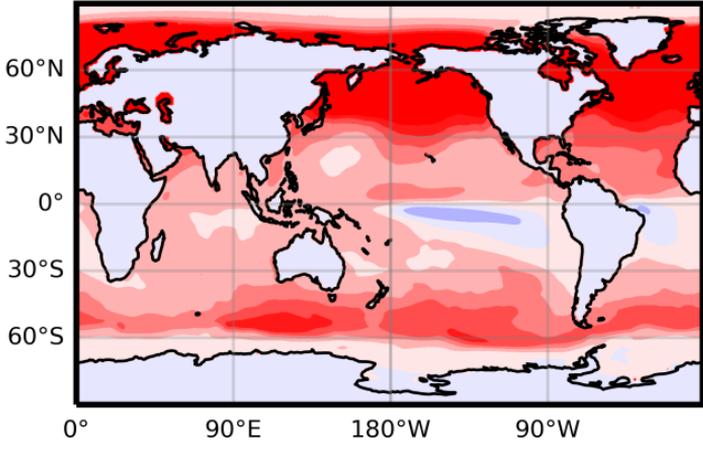


Figure 7.

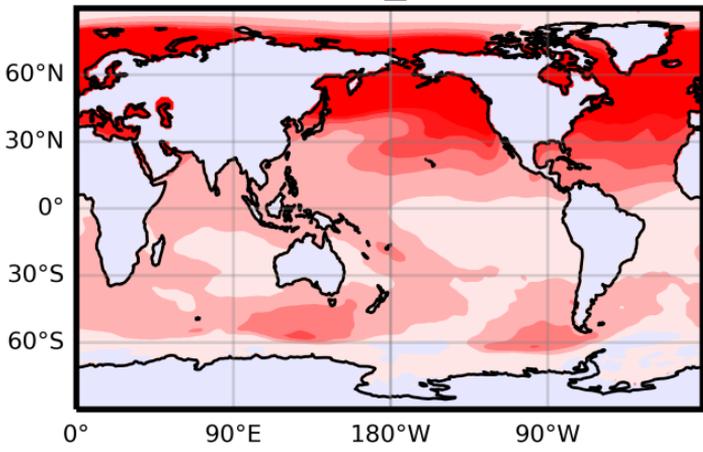
RESP_OGCM



RESP_EMOM



RESP_MLM



RESP_SOM

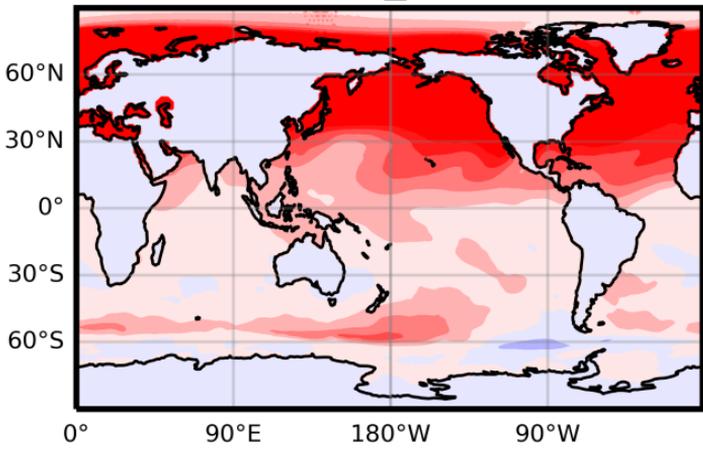


Figure 8.

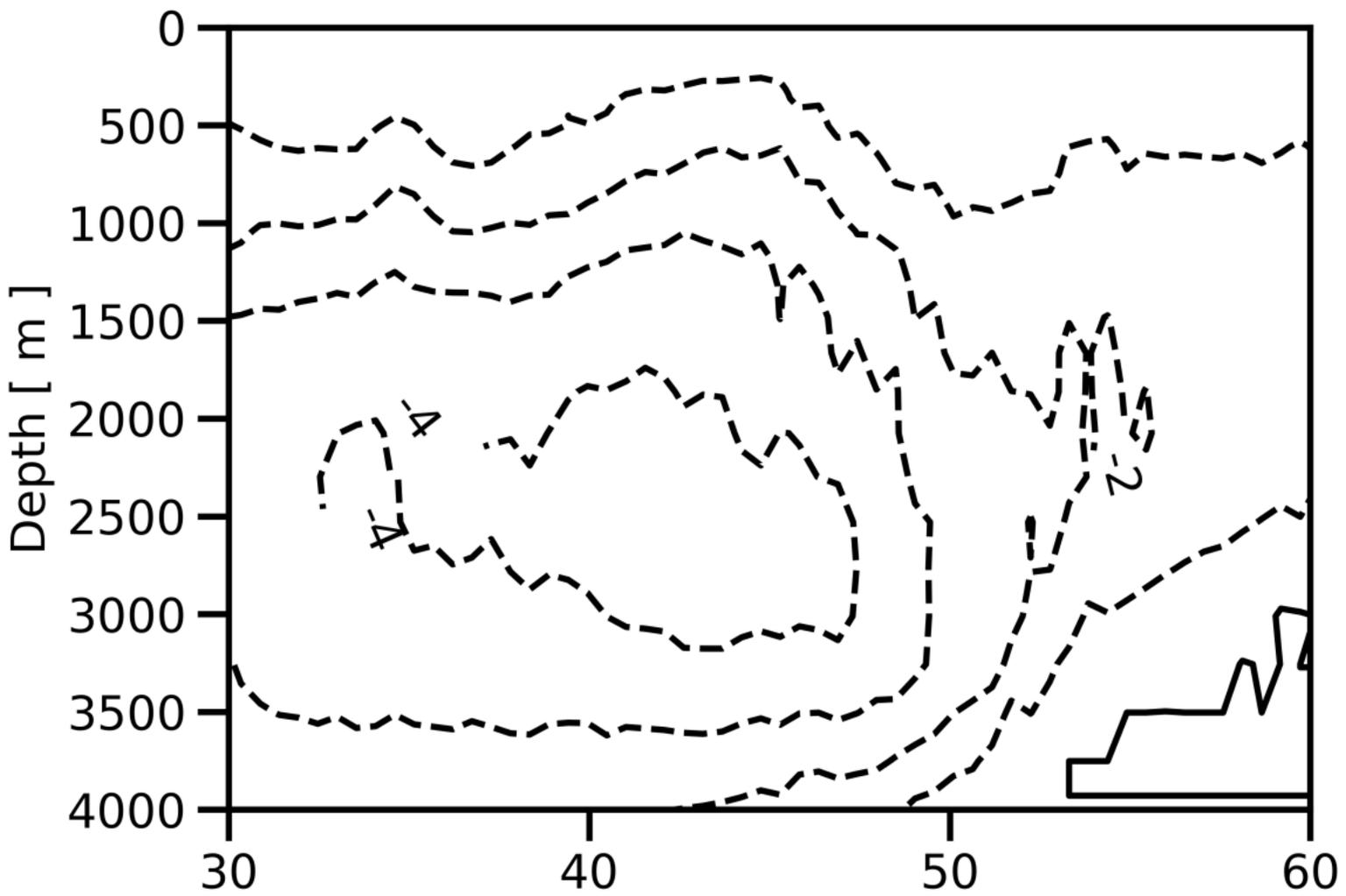


Figure 9.

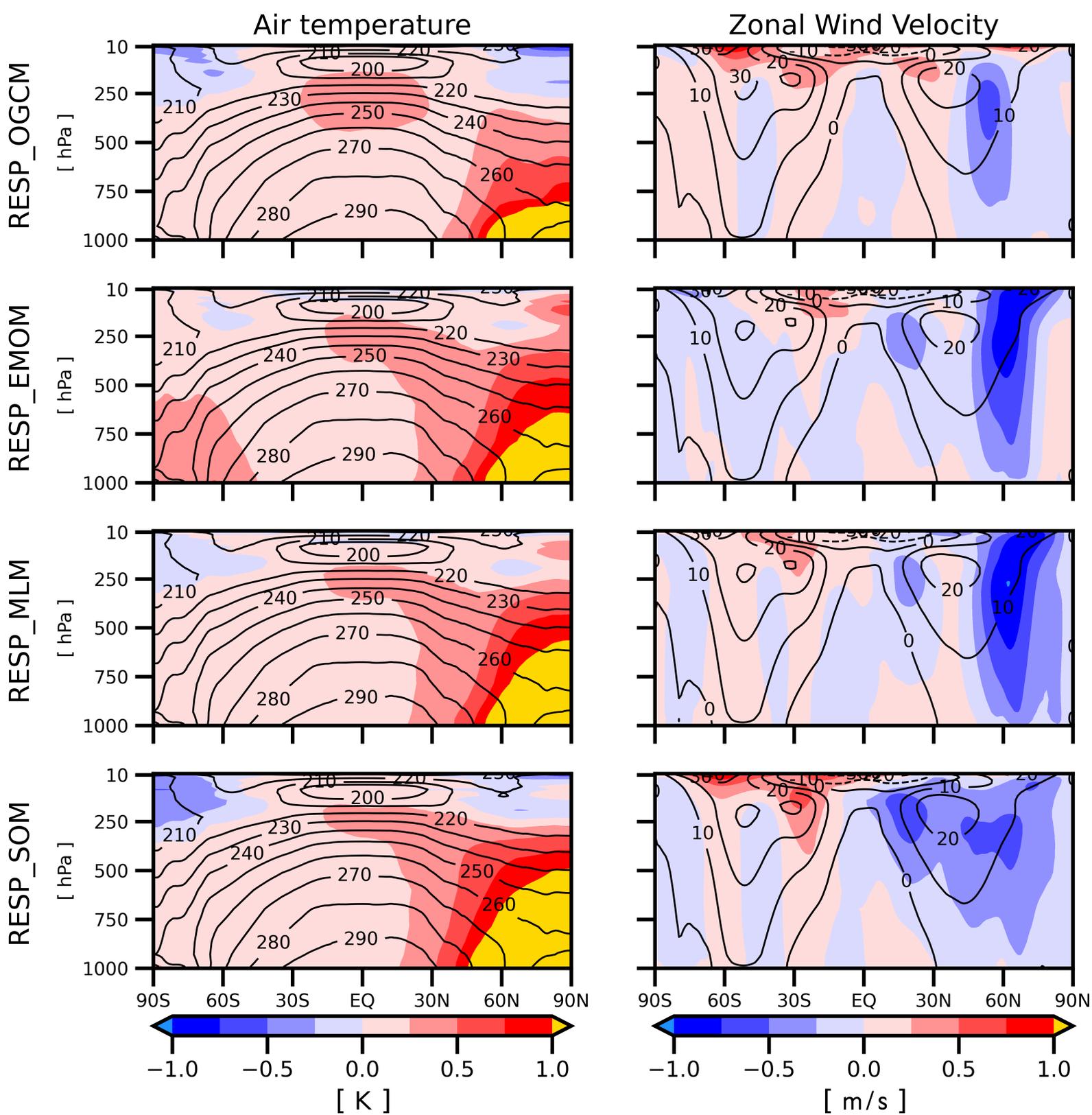
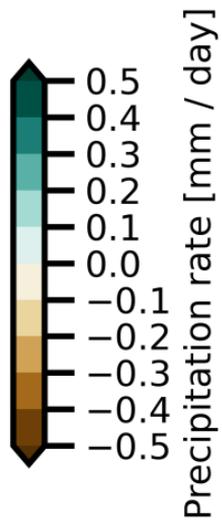
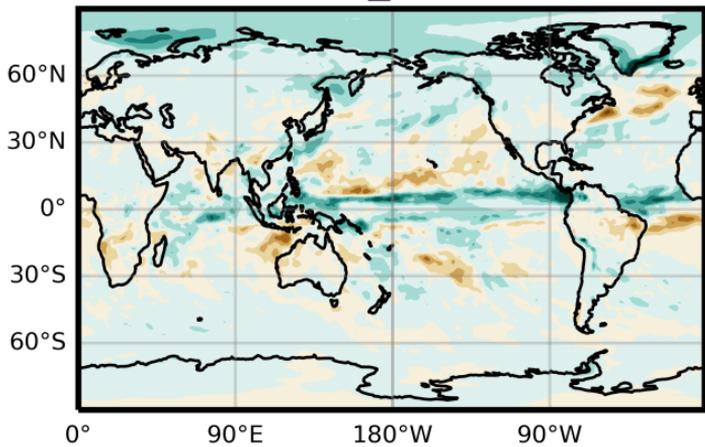
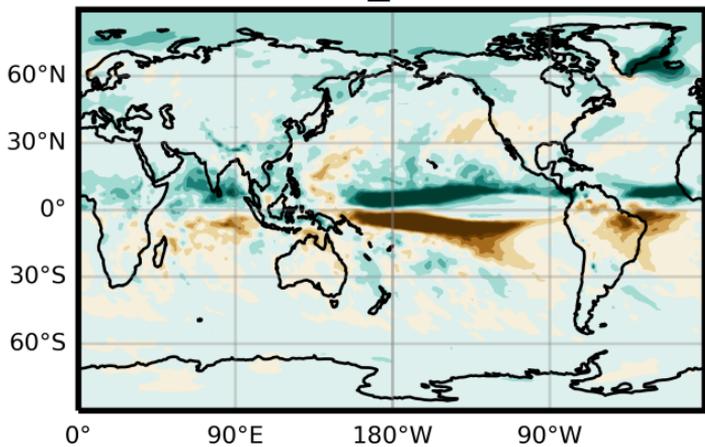


Figure 10.

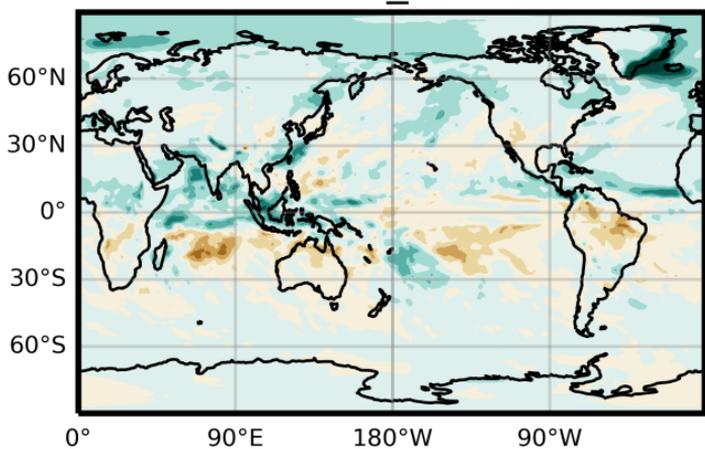
RESP_OGCM



RESP_EMOM



RESP_MLM



RESP_SOM

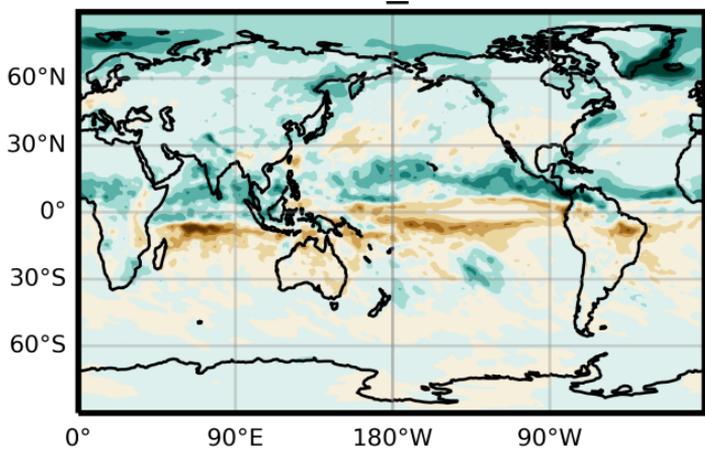


Figure 11.

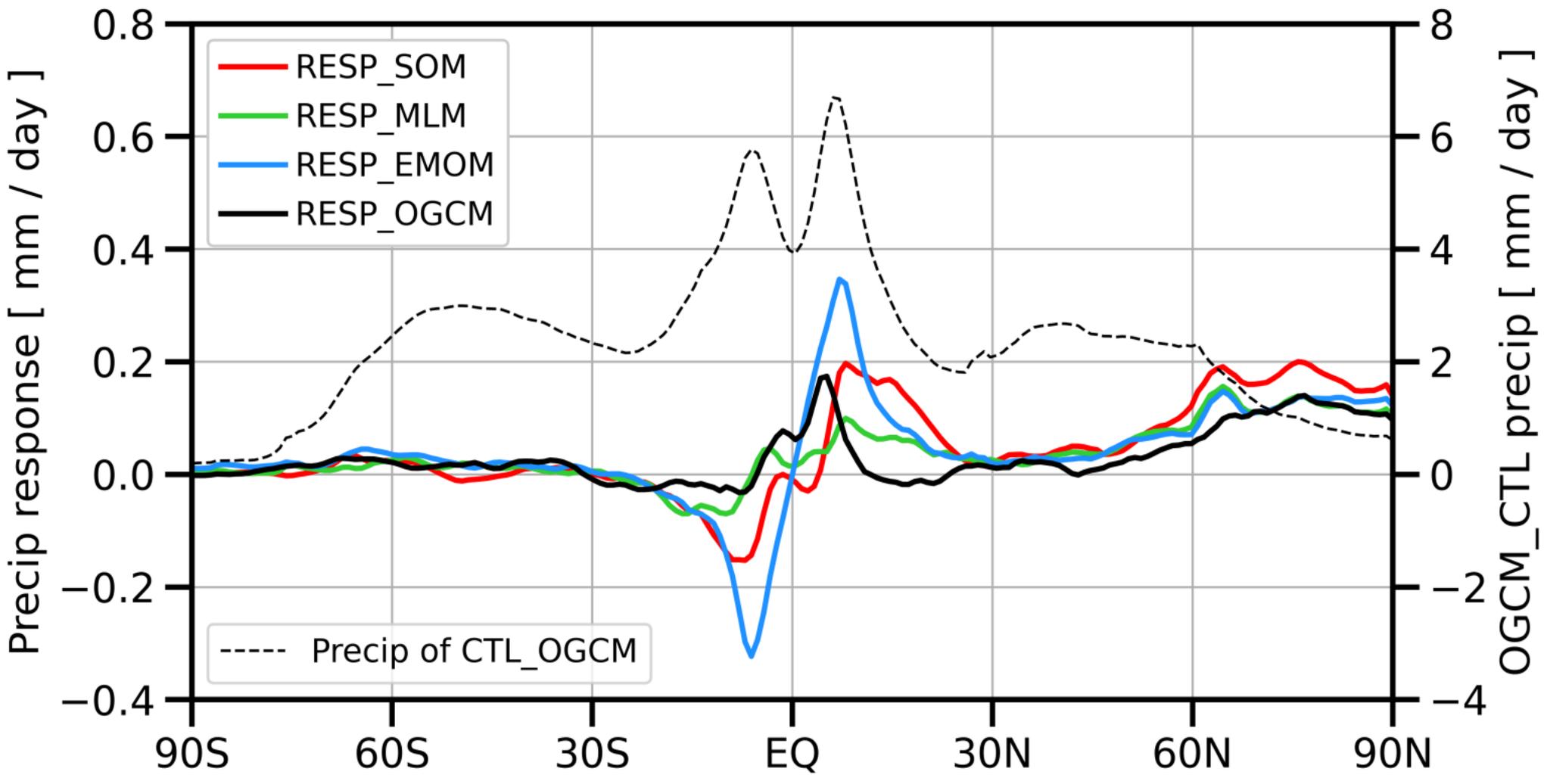


Figure 12.

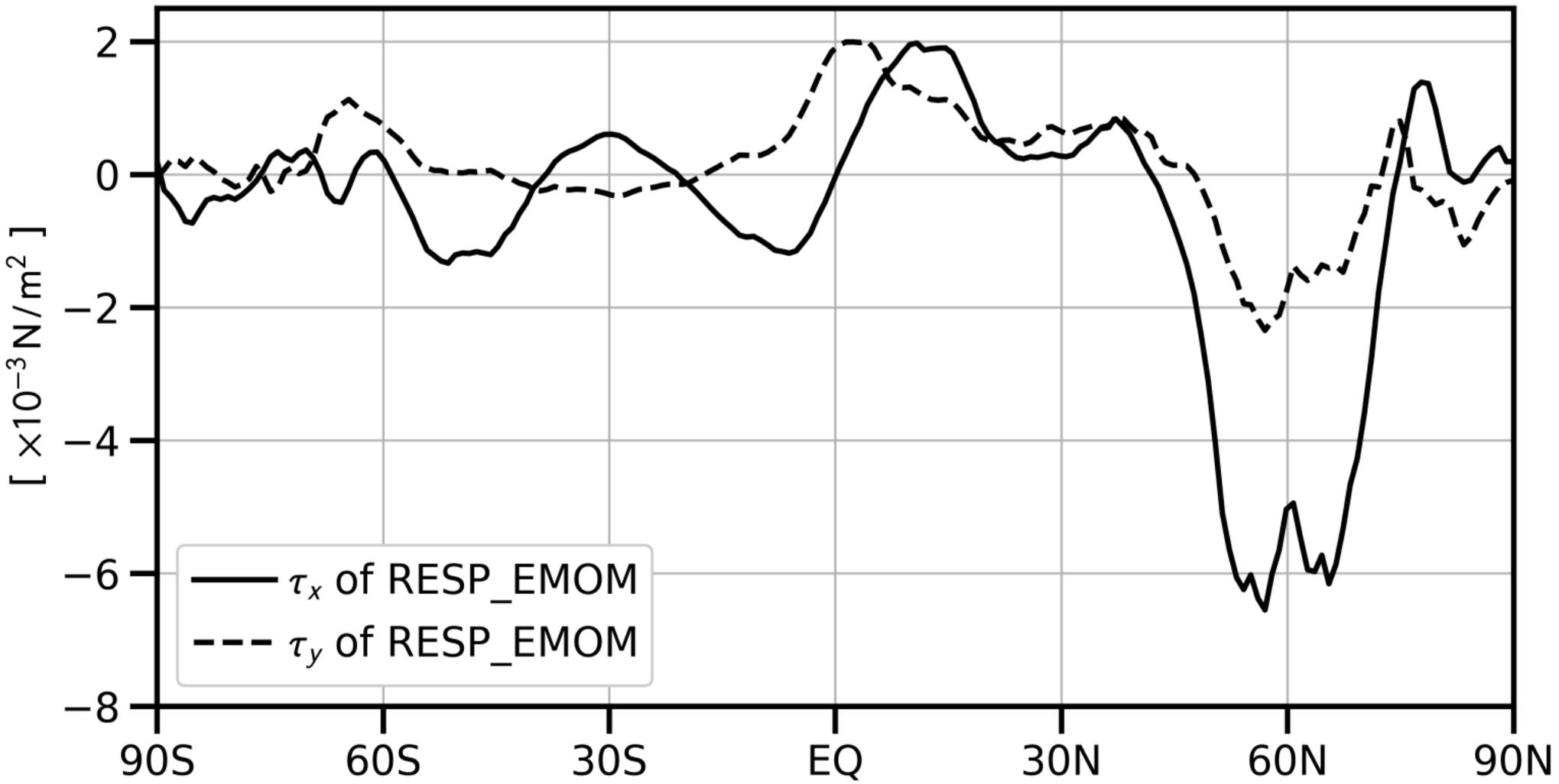
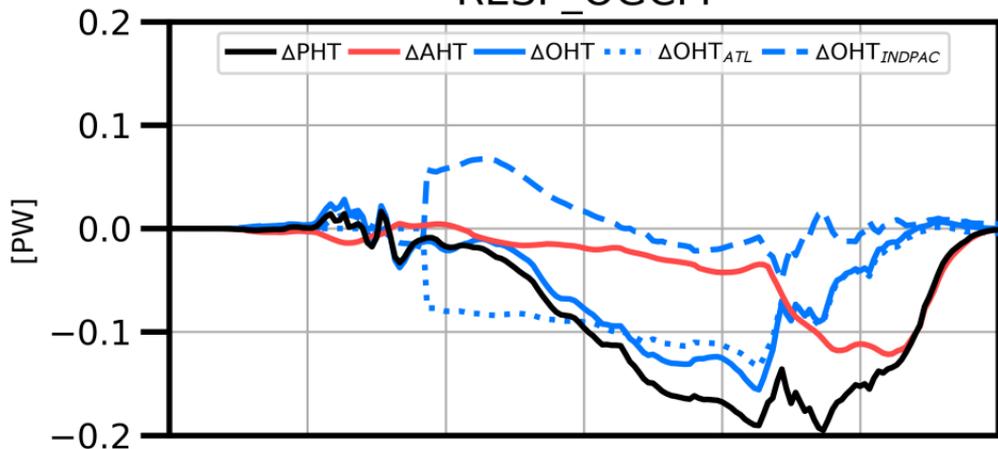
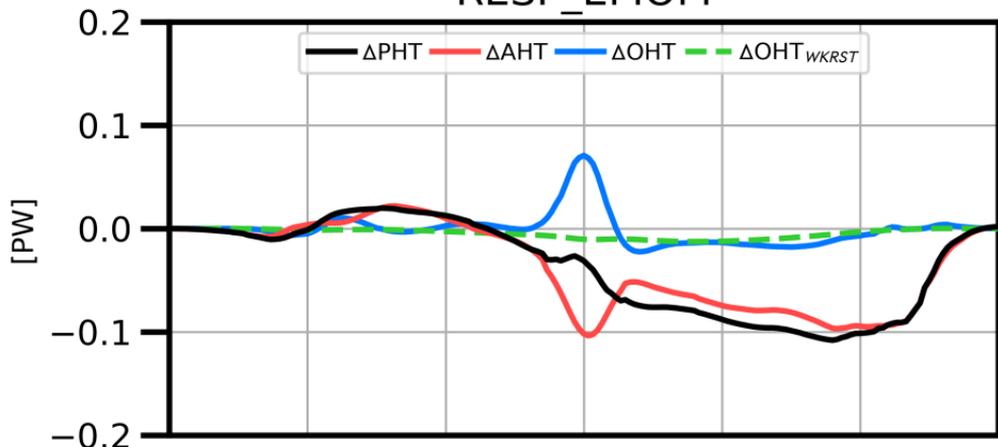


Figure 13.

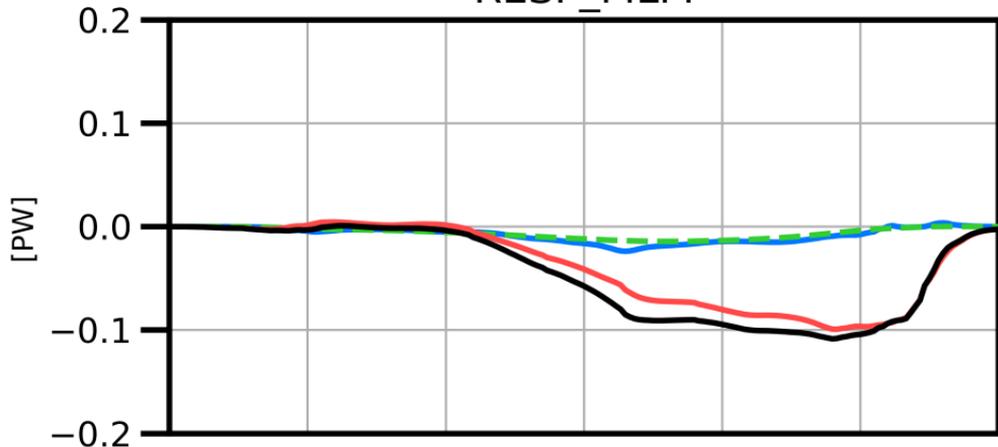
RESP_OGCM



RESP_EMOM



RESP_MLM



RESP_SOM

