A Partial Coupling Method to Isolate the Roles of the Atmosphere and Ocean in Coupled Climate Simulations

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Abstract

This study describes the formulation and application of a partial coupling method that disentangles the coupling between the atmosphere and ocean and isolates the atmosphere- and ocean-driven components of the coupled climate interactions. In contrast to strategies using stand-alone simulations with prescribed atmosphere or ocean states, the climate components in the partially coupled method remain coupled, but the impact of ocean circulation changes is removed from the air-sea interaction using temperature-like tracers. The partially coupled simulation thereby suppresses the ocean-driven interaction and isolates an atmosphere-driven interaction only. The ocean-driven component can be inferred by comparing climate response in the partially coupled simulation with that of a standard fully coupled one. The partial coupling approach is applied to decompose the fully coupled climate response to CO2 quadrupling into atmosphere- and ocean-driven components. The linearity of the decomposition is validated by simulating the ocean-driven response using another complimentary partially coupled simulation forced only with the atmosphere-driven anomalous surface fluxes. A comparison of the two partially coupled simulations with the fully coupled simulation indicates that the sum of the atmosphere- and ocean-driven components accurately describes the fully coupled response. 

Plain Language Summary

Climate change and variability include components that are produced by both atmospheric and ocean changes. This study describes a method that disentangles the interaction between the atmosphere and ocean and separates out the components of climate change that are caused by atmospheric and oceanic changes. Two climate experiments with quadrupled CO2 concentration are used to demonstrate this method. In one experiment the component of the ocean temperature change that is produced by ocean circulation changes is prevented from interacting with the atmosphere at the surface, so that the climate change in this experiment is caused by the atmospheric CO2 increase alone. Typical coupling strategies allow both the interactions caused by atmospheric CO2 increases and ocean circulation changes to occur. The difference between the two experiments isolate the climate change component that is caused by ocean circulation changes. An additional experiment was used to reproduce the climate change component caused by the ocean circulation changes, in order to show that the two components sum up to the total climate change when all interactions caused by the atmosphere and ocean are allowed.

1. Introduction

Temporal variation in the earth system (internal variability and responses to external forcing) is intimately tied to the interactions between its dynamic components, that is, the atmosphere and ocean. The exchange of energy between the atmosphere and ocean occurs at the air-sea interface through surface heat, salinity, and momentum fluxes. The exchange allows for a two-way interaction, whereby the atmosphere drives the ocean and the ocean also drives the atmosphere. However, this tangled two-way interaction also makes it difficult to isolate the individual roles of the atmosphere and ocean in earth system variability and response to external forcing with their associated timescales. Isolating the individual roles of the atmosphere and ocean is helpful to understand the mechanisms of climate variability and change and improve climate predictions or diagnose climate model simulations.
The traditional approach to evaluating the individual roles of the atmosphere and ocean involves the use of “stand-alone experiments,” frequently called “atmosphere-only,” “ocean-only,” and “slab ocean model” experiments. These uncoupled or constrained atmosphere or ocean model experiments use prescribed SSTs, surface fluxes, or ocean mixed layer heat capacity and heat convergence respectively, to represent ways that one component (ocean, atmosphere, etc.) can influence the energy exchanges at the component interface when the other component of the system has a limited capacity to respond. However, this treatment of the boundary interactions often limits the attribution of the role of the atmosphere or ocean in stand-alone experiments. Prescribing boundary interactions can obscure the interactions across timescales between the atmosphere and ocean and can cause overestimation or underestimation of the roles of the atmosphere or ocean, or worse, can confound roles of each component in fully coupled simulations.

For example, prescribed SSTs and their anomalies in atmosphere-only simulations are often attributed to the ocean, but the SSTs are often derived from a fully coupled system with a two-way interaction between the atmosphere and ocean. Prescribed SST anomalies are therefore partly driven by the atmosphere, and attributing the anomaly cause to the ocean alone can produce a misleading understanding of the atmosphere’s role. Also, prescribed SSTs do not allow for any response from the underlying ocean and thus can produce inconsistent or unrealistic energy fluxes at the air-sea interface (Bretherton & Battisti, 2000; Deser & Phillips, 2009; Deser & Timlin, 1997; Dong et al., 2017; Folland et al., 1998; He & Soden, 2016; He et al., 2017, 2018; Hirons et al., 2018; Kumar & Hoerling, 1998). Slab ocean experiments overcome the problem of inconsistent or implied energy fluxes in atmosphere-only experiments by coupling the atmosphere model to an ocean mixed layer with prescribed ocean heat convergence. This allows for SSTs to respond to atmospheric changes but without the two-way interactions present in a fully coupled system which allows the ocean also to drive atmospheric changes through its circulation response (Barsugli & Battisti, 1998; Battisti et al., 1995; Bitz et al., 2012). Nonetheless, due to their prescribed ocean heat convergence and mixed layer depth, slab ocean experiments do not allow for changes of the ocean heat convergence and mixed layer depth that accompany anomalous surface heat fluxes or momentum transfer at the air-sea interface (Larson et al., 2018). This limitation could also lead to a wrong attribution of the role of the atmosphere in regions where ocean heat convergence can be large (Clement et al., 2015; Garuba, Lu, Singh, et al., 2018; Zhang, 2017).

Ocean-only simulations are likewise produced by prescribing surface fluxes or by restoring ocean surface boundary to prescribed conditions or by using a combination of both methods (mixed boundary conditions, as in Griffies et al., 2009; Haney, 1971; Rahmstorf & Willebrand, 1999). Prescribed surface fluxes or restoring boundary conditions are also often derived from fully coupled system with a two-way interaction, driven by both atmosphere and ocean; therefore, their use can lead to a double counting or wrong attribution of the ocean’s role. Prescribed surface fluxes also do not allow for the feedback between anomalous surface fluxes and the ocean surface response which damps anomalous surface fluxes over time in fully coupled air-sea interaction. The lack of this feedback can cause large drifts in ocean-only experiments. Restoring boundary formulations overcomes this limitation and allows for a feedback between anomalous surface fluxes and the ocean surface temperature response, but the solutions are dependent on the restoration time scale (or coupling strength) chosen or the boundary conditions that are prescribed. An unrealistic choice of restoring timescale or prescribed boundary conditions that are derived from a climate state that differs significantly from the relevant ocean model state can also produce very different surface flux response (Cai & Chu, 1996; Chu et al., 1998; Frankignoul et al., 1998; Killworth et al., 2000; Large et al., 1997; Marotzke, 1994; Pierce, 1996).

Given the limitations of stand-alone experiments, we describe here an alternate partial coupling method (both details about the formulation and validation) that allows for the isolation of the individual roles of the atmosphere and ocean within a coupled model framework. The atmosphere and ocean components remain coupled, but the impact of ocean circulation changes is removed from the air-sea interactions. The partially coupled experiment thereby suppresses ocean-driven surface fluxes and simulates atmosphere-driven anomalous surface fluxes and response alone. The ocean-driven contributions can be recovered by taking the difference between the partially coupled experiment and a standard fully coupled experiment with a two-way interaction. This isolation of the atmosphere-driven anomalous surface heat fluxes employs the tracer decomposition methods introduced in Banks and Gregory (2006) and Xie and Vallis (2012), where
additional temperature-like passive tracers were added to an ocean model to isolate anomalous ocean temperature components that were driven by anomalous surface heat fluxes and ocean circulation changes respectively. Here in the partially coupled experiment, one of the tracers is also used in the coupling with the atmosphere in order to remove the impact of the ocean circulation changes.

The partial coupling method is similar in spirit to some earlier strategies for isolating the roles of atmosphere and ocean in coupled air-sea interactions. Winton et al. (2013) and Garuba and Klinger (2016, 2018) all removed the impact of ocean circulation changes from their surface interactions in order to suppress ocean-driven surface fluxes and assess impacts of the atmosphere-driven ones. Ocean circulation changes were prevented in their global warming experiment in the Winton et al. (2018b) study, while ocean circulation-driven SST changes were removed from the restoring boundary formulation in the ocean-only study of Garuba and Klinger (2016, 2018). Similarly, in slab model studies, ocean circulation changes and their impact were disabled using an ocean mixed layer with prescribed heat transports and mixed layer depths. The studies of Bouttes et al. (2014) and Gregory et al. (2016) also use a different partial coupling approach to simulate rather than suppress ocean-driven anomalous surface fluxes. However, these earlier methods used imposed boundary conditions that are susceptible to the previously mentioned double or under counting problem. The partially coupled framework described here overcomes this limitation and also allows for the isolation of both atmosphere- and ocean-driven surface heat flux anomalies as well as the atmosphere and ocean responses to these changes (provided that the fully coupled response is linear).

We discuss the partitioning of the air-sea interaction into atmosphere- and ocean-driven components and the partially coupled experiments used to isolate and validate the decomposition in section 2. The method is then applied to isolate the atmosphere- and ocean-driven components of the climate response to CO2 quadrupling, by comparing partially and fully coupled CO2 quadrupling experiments. In section 3, features of the atmosphere- and ocean-driven responses are compared and discussed. The atmosphere- and ocean-driven responses are then shown to add up to the fully coupled response by using an independent simulation of the ocean-driven response. The atmosphere- and ocean-driven responses are shown to be consistent with atmospheric perturbations and the ocean circulation change producing them. In section 4, we discuss and contrast our conclusions to those produced using the other methods mentioned in this study.

2. Method Formulation

2.1. Partitioning the Fully Coupled Air-Sea Interaction

Air-sea flux anomalies can originate from either the atmosphere or ocean component of the climate system. Atmospheric temperature, hydrologic, or circulation changes can cause anomalous heat, salinity, and momentum fluxes at the air-sea interface that can drive an ocean response. The ocean also produces surface flux changes (heat, salinity, and momentum) mainly through its sea surface temperature (SST) changes and can in turn drive an atmospheric response. SST changes can result from any of the anomalous surface flux components arising in the atmosphere, since they are primarily caused by anomalous surface heat fluxes and ocean circulation changes, and ocean circulation changes can be produced by any of the anomalous surface flux components. The ocean can thus in turn drive additional surface flux changes in response to any atmosphere-driven surface flux perturbation.

The partitioning of this two-way or fully coupled air-sea interaction into components that are driven by the atmosphere and ocean may be described conceptually by using an idealized equation for the air-sea heat fluxes:

\[ Q = \alpha(T_A - T_O)l_s \]
\[ Q' = \alpha'\left(T'_A - T'_O\right)l_s \]

Here, \( Q \) is the surface heat flux, \( T_A \) is the air temperature, \( T_O \) is the ocean temperature, \( \alpha \) is the coupling coefficient (varying in space and time) representing the strength or timescale of the surface coupling between the atmosphere and ocean and depends on the surface wind strength, and \( l_s \) denotes the surface values of the variables. These variables include a baseline state and anomalous component denoted by overbars and primes respectively (i.e., \( Q = Q + Q' \)). The baseline may be defined as the climatological base state or unperturbed base state, depending on the anomaly time scale of interest. Equation 1 is shown to be a good
The net surface heat temperature perturbations can be partitioned into components developing originally within the ocean or atmosphere. Differentiating the secondary surface temperature responses from the original surface temperature perturbation producing the anomalous surface temperature changes will then also cause further changes in the anomalous surface fluxes. These additional surface temperature changes will then also cause further changes in the anomalous surface fluxes. However, surface temperature responses to the initial anomalous surface heat fluxes are functions of changes in the air-sea interaction or the driving ocean surface temperature component is removed such that the ocean acts to only damp. Note that the net surface temperature changes of driving climate component in these one-way interactions also include damping responses (i.e., $T_{AQ}$ and $T_{OO}$). The inclusion of the damping

approximation of the bulk formula that is actually used in computing the turbulent surface flux exchange in fully coupled models and is also used for the restoring boundary formulation for ocean-only experiments (Haney, 1971; Rahmstorf & Willebrand, 1995), where $T_A$ is an apparent surface air temperature. Though it is primarily applied to the turbulent surface flux components (momentum, sensible, and latent heat fluxes), this approximation can be extended to the other surface flux components (salinity and radiative heat fluxes) for the purpose of partitioning the air-sea interaction, as these are also functions of the turbulent fluxes (through evaporation and clouds) and indirect functions of the surface temperature changes (Rivin & Tziperman, 1997).

Equation 1 suggests that anomalous surface heat fluxes are functions of changes in the air-sea temperature difference. A perturbation in one component's (ocean or atmosphere) surface temperature will cause anomalous surface heat fluxes that will induce additional surface temperature changes in both components. These additional surface temperature changes will then also cause further changes in the anomalous surface fluxes. However, surface temperature responses to the initial anomalous surface heat fluxes generally act to damp the flux anomaly by reducing the temperature difference. The secondary surface temperature response thus provide a negative feedback that damp rather than drive further anomalous surface fluxes over time. Differentiating the secondary surface temperature responses from the original surface temperature perturbation that produce anomalous surface fluxes is therefore useful for partitioning the anomalous surface fluxes.

The net surface heat flux ($Q'$) can then be partitioned into “atmosphere-driven” ($Q'_{A}$) and “ocean-driven” ($Q'_{O}$) components based on where the surface temperature perturbation producing the flux originated. Surface temperature perturbations can be partitioned into components developing originally within the ocean or atmosphere component ($T'_{AA}$ and $T'_{OO}$, respectively) or as a response to an exchange of heat across the air-sea interface ($T'_{AQ}$ and $T'_{OO}$). It is also useful to track whether the temperature responses to $Q'$ arose from the $Q'_{A}$ or $Q'_{O}$ component. So we further decompose $T'_{AQ}$ into $T'_{AQ_a}$ and $T'_{AQ_o}$, and similarly $T'_{OO}$ into $T'_{OO_a}$ and $T'_{OO_o}$. We can then rewrite Equation 1 relating ocean and atmosphere temperature perturbations and heat fluxes into a few forms useful in describing driving and damping terms:

$$Q' = \alpha((T'_{AA} + T'_{AQ}) - (T'_{OO} + T'_{OO}'))|_s$$

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$$= \alpha((T'_{AA} + T'_{AQ_a} + T'_{AQ_o}) - (T'_{OO} + T'_{OO_a} + T'_{OO_o}'))|_s$$

Equation 2 can then be rearranged into the atmosphere- and ocean-driven components, that is,

$$Q' = \alpha((T'_{AA} + T'_{AQ_a}) - T'_{OO_a}) + (T'_{AQ_o} - (T'_{OO} + T'_{OO_o}))|_s$$

$$Q'_{A} = \alpha((T'_{AA} + T'_{AQ_a}) - T'_{OO_a})|_s$$

$$Q'_{O} = \alpha(T'_{AQ_o} - (T'_{OO} + T'_{OO_o}))|_s$$

The rearrangement of the terms in Equation 3 suggests that the atmosphere-driven surface heat fluxes ($Q'_{A}$) are the surface fluxes occurring if only the driving atmosphere surface temperature ($T'_{AA}$) is allowed to drive the air-sea interaction or the driving ocean surface temperature component is removed such that the ocean surface temperature response ($T'_{OO_a}$) acts only to damp air-sea interaction. In contrast, the ocean-driven components ($Q'_{O}$) are the anomalous surface fluxes occurring if the air-sea interaction is driven only by driving ocean temperature ($T'_{OO}$), and the atmosphere surface temperature response ($T'_{AQ_o}$) acts only to damp the air-sea interaction. $Q'_{A}$ or $Q'_{O}$ would be produced in a simulation that supports a one-way air-sea interaction, driven only by one climate component (atmosphere or ocean), while the other climate component's response acts to only damp. Note that the net surface temperature changes of driving climate component in these one-way interactions also include damping responses (i.e., $T_{AQ}$ and $T_{OO}$). The inclusion of the damping
surface temperature response of the nondriving climate component in this partitioning is the key to providing the negative feedback on these anomalous one-way air-sea interactions that prevents a drift of the experiments in which they are implemented.

2.2. Ocean Temperature Decomposition

In order to realize the hypothetical atmosphere- or ocean-driven one-way air-sea interactions discussed following Equation 3, it is sufficient to isolate the driving and damping ocean temperature components. Since the ocean interacts primarily through its SSTs, removing the driving ocean temperature component from the air-sea interaction or driving the anomalous air-sea interaction with it would produce atmosphere- and ocean-driven surface fluxes, respectively.

It is useful then to identify the driving and damping ocean temperature responses to an atmospheric forcing in a standard fully coupled interaction. To do this, consider a simplified anomalous ocean temperature tendency equation (where heat capacity and density are ignored because they are not germane to the decomposition):

$$\frac{\partial T'_o}{\partial t} = Q' - v' \cdot \nabla T'_o - v \cdot \nabla T'_o$$

Equation 4 is derived by subtracting the ocean temperature tendency equation for a baseline state from that of a time varying or perturbed state (see Appendix A). Variables without overbars or primes indicate the total (baseline state and anomaly, that is, $T'_o = T'_o + T'_o$, $v = v + v'$; $Q = Q + Q'$). Here, Equation 4 applies to the full 3-D ocean temperature field; $v$ is the 3-D ocean circulation strength, and the transport terms that
represent all transport processes (including resolved advection and subgrid mixing and diffusion). Equation 4 is a general equation of the ocean temperature change and can apply to any ocean temperature layer. \( Q' \) is nonzero only at the surface and is produced by the air-sea interaction described by Equations 2 or 3.

According to (4), ocean temperature changes may result from both anomalous surface heat fluxes, \( Q' \), and ocean circulation changes, \( \nu' \), shown by the first two source terms. The third term does not drive ocean temperature changes but can determine how quickly it responds to surface heat flux and ocean circulation changes. As discussed in section 2.1, if \( T'_o \) is expanded into \( T'_{oo} \) and \( T'_{oo} \), and \( T'_{oo} \) is defined to be the part of \( T'_o \) that is the response to \( Q' \) then one can collect all terms involving \( T'_{oo} \) from (4) to form a new equation, and the remaining terms are used to describe the evolution of \( T'_{oo} \). Equation 4 may then be split into a sum of two parts based on these definitions:

\[
\frac{\partial T'_{oo}}{\partial t} = Q' - \nu \cdot \nabla T'_{oo} \tag{5}
\]

\[
\frac{\partial T'_{oo}}{\partial t} = -\nu' \cdot \nabla T'_o - \nu \cdot \nabla T'_{oo} \tag{6}
\]

This partitioning of the anomalous ocean temperature is also known as the surface- and circulation-driven partitioning of the ocean temperature. \( T'_{oo} \) is the surface-driven component produced by the anomalous surface heat fluxes \( Q' \); it is also the damping component of the fully coupled air-sea interaction which damps both atmosphere- and ocean-driven components of \( Q' \). While \( T'_{oo} \) is the circulation-driven component and driving surface temperature component that produces the ocean-driven component of the air-sea interaction \( (Q'_o) \). \( T'_{oo} \) has also been referred to as a "redistributive" temperature anomaly because it is caused by the rearrangement of the baseline ocean temperature gradients by circulation changes; \( T'_{oo} \) therefore cannot cause a change in the global ocean heat content change but can contribute to regional ocean heat content changes. \( T'_{oo} \) can result from ocean circulation changes caused by an internal perturbation or variability in the ocean or produced as a response to atmosphere-driven anomalous surface fluxes. In the latter case, \( T'_{oo} \) produces further surface fluxes changes rather than damp the atmosphere-driven surface fluxes produced by an atmospheric perturbation. Since most perturbations to the climate system originate in the atmosphere, the latter case is what is described in this study.

The damping (surface-driven) and driving (circulation-driven) anomalous ocean temperature components can be isolated following the tracer decomposition method of Banks and Gregory (2006) and Xie and Vallis (2012). Two additional temperature-like tracers are introduced in the ocean model simulation that evolve according to a generic tracer equation:

\[
\frac{\partial P_i}{\partial t} = Q_i - \nu \cdot \nabla P_i \tag{7}
\]

Initial conditions and forcing terms \((Q_i)\) can be chosen for the tracers \((P_i)\) to provide information in a simulation (described in following paragraphs). In a fully coupled simulation these tracers serve only to help in diagnostics; they do not change ocean density or currents or participate in the air-sea interaction.

The first tracer, \( P_1 \), is designed to track the surface-driven or damping ocean temperature anomalies \((T'_{oo})\). \( P_1 \) is set to zero everywhere at the initialization of the experiment and forced at the surface with anomalous surface heat fluxes \( Q_1 \), constructed by subtracting a baseline surface heat flux from the coupled surface heat flux exchanged in the experiment \((i.e., Q_1 = Q' = Q - Q)\). Thus, \( P_1 \) integrated using Equation 7 provides a solution equivalent to Equation 5. The second tracer, \( P_2 \), is designed to represent the evolution of the baseline ocean temperature field \((T_o)\) produced by the baseline surface heat flux \((Q)\); \( P_2 \) is thus set equal to \( T_o \) at initialization and also forced with \( Q \) \((i.e., Q_2 = Q)\). However, over time, \( P_2 \) is expected to deviate from the baseline temperature field evolution, when the ocean circulation strength in the experiment with tracer \( P_2 \) and different from the baseline circulation \((\nu \neq \nu)\). It can be shown that the difference between the tendency equations for tracer \( P_2 \) and \( T_o \) is equal to the circulation-driven or damping ocean temperature anomaly Equation 6 (see the Appendix B for derivation of tracer equations). \( T'_{oo} \) is thus obtained as the difference
In order to realize this one-way air-sea interaction, additional tracers $P_1$ and $P_2$ that evolve according to Equation 7 are again used to isolate and diagnose the surface- and circulation-driven anomalous ocean temperature components, but this time the tracer $P_1$ actually participates in the calculation of the surface flux exchange. As previously described for the tracers in the fully coupled experiment, initial values for $P_1$ and $P_2$ are set respectively to zero and $T_{o,0}$ at initialization, and the tracers are forced with the anomalous surface heat fluxes computed online and a baseline flux respectively (i.e., $Q_1 = Q_{cp} - Q$ and $Q_2 = Q$). $T_{o,0}$ and $Q$ are similarly taken from the baseline experiment used for forcing the tracers in the fully coupled experiment. The ocean temperature used in computing the surface flux exchange is constructed by adding tracer $P_1$ to

| Experiment | Tracer | Initialization | Surface source derivation | Temperature equivalence and coupling |
|------------|--------|----------------|--------------------------|-------------------------------------|
| Fully coupled | $P_1$ | $P_1|_{t=0} = 0$ | Computed ($Q_1 = Q = Q_{cp} - Q$) | $P_1 = T_{o,0}$ |
| Partially coupled (atmosphere-driven) | $P_2$ | $P_2|_{t=0} = T_0$; Prescribed ($Q_2 = Q$) | $P_2 = T_0 + T_{o,0}$; |
| Partially coupled (ocean-driven) | $P_1$ | $P_1|_{t=0} = 0$ | Computed ($Q_1 = Q = Q_{cp} - Q$) | $P_1 = T_{o,0}$; |
| Partially coupled (ocean-driven) | $P_2$ | $P_2|_{t=0} = T_0$; Prescribed ($Q_2 = Q$) | $P_2 = T_0 + T_{o,0}$; |

Note. $Q$ indicates a surface flux, $T$ a temperature-like variable. The subscript $cpl$ is used to indicate the heat flux that is exchanged between the ocean and atmosphere $Q_{cpl}$, and the ocean temperature that is used for calculating for the tracers $P_1$ and $P_2$, respectively. The tilde above $Q_1$ is used to indicate a prescribed surface flux anomaly in the ocean-driven partially coupled experiment.

between tracer $P_2$ and $T_0$ (i.e., $P_2 = T + T_{o,0}$), and the sum of the two tracers is equivalent to the actual ocean temperature ($T_0 = P_1 + P_2$, see also Figure B1). $Q$ and $T_0$ are supplied as input fields for forcing the tracers and are derived from a baseline experiment. These relationships are summarized in Table 1.

### 2.3. Atmosphere-Driven Anomalous Surface Flux Simulation

It is useful to begin this section by introducing a variable we will call $Q_{cpl}$, which is flux computed in the model component called coupler and is the flux exchanged between atmosphere and ocean models. In a fully coupled experiment, the flux calculation uses the actual ocean temperature ($T_o$) and $Q_{cpl}$ is equivalent to $Q$ in Equation 1 that contains the baseline and all the anomalous surface flux terms. In the partially coupled experiments described here and in this section, a surrogate ocean temperature ($T_{cpl}$) is constructed and used in the flux calculation instead of the model’s standard ocean temperature, that heat flux is then exchanged between the ocean and atmospheric models and drives them. $T_{cpl}$ makes use of a baseline ocean temperature and one anomalous ocean temperature component isolated by the evolving tracer fields that are carefully chosen to make it possible to suppress various terms in Equation 3. The use of the tracers in constructing $T_{cpl}$ and $Q_{cpl}$ are described in this and the following sections.

As defined in Equation 3, atmosphere-driven surface fluxes are the anomalous surface fluxes produced if the circulation-driven anomalous ocean temperature component is excluded from participating in the air-sea interaction. Therefore, in an atmosphere-driven partially coupled experiment (with a perturbation originating in the atmosphere model), only the damping or surface-driven ocean temperature response is allowed to participate in the air-sea interaction, so that its anomalous surface fluxes originate only from the atmosphere (i.e., $Q'_{a}$), and its damping ocean temperature anomaly is equivalent to $T'_{o,0}$ as defined in Equation 3 (Figure 1b). The ocean-driven anomalous surface flux component ($Q'_{o}$) may then be inferred from this partially coupled experiment by taking the difference between its anomalous surface fluxes and that of a fully coupled experiment with similar initialization and perturbation (i.e., $Q'_{o} = Q' - Q'_{a}$).

In order to realize this one-way air-sea interaction, additional tracers $P_1$ and $P_2$ that evolve according to Equation 7 are again used to isolate and diagnose the surface- and circulation-driven anomalous ocean temperature components, but this time the tracer $P_1$ actually participates in the calculation of the surface flux exchange. As previously described for the tracers in the fully coupled experiment, initial values for $P_1$ and $P_2$ are set respectively to zero and $T_{o,0}$ at initialization, and the tracers are forced with the anomalous surface heat fluxes computed online and a baseline flux respectively (i.e., $Q_1 = Q_{cp} - Q$ and $Q_2 = Q$). $T_{o,0}$ and $Q$ are similarly taken from the baseline experiment used for forcing the tracers in the fully coupled experiment. The ocean temperature used in computing the surface flux exchange is constructed by adding tracer $P_1$ to
the baseline ocean temperature to (i.e., $T_{\text{cpl}} = T_0 + P_1|_t \equiv T_0 + T_{\text{OO}}|_t$). The form of tracer equations and fluxes can be compared with the fully coupled configuration by referencing Table 1.

Since $P_1$ is zero at the initial time, the surface fluxes are initially computed with the baseline ocean temperature alone, so that the anomalous surface fluxes are produced only by the atmospheric surface temperature perturbation (i.e., $Q_{\text{cpl}} = Q + Q'_{A}$). $P_1$ is then forced with the atmosphere-driven anomalous surface fluxes computed online in the experiment (i.e., $Q_1 = Q_{\text{cpl}} - Q = Q'_{A}$). $P_1$ thus becomes nonzero over time ($T_{\text{OO}}$), and also participates in the air-sea interaction but only to damp $Q'_{A}$, according to Equation 3. Due to the differences between the anomalous surface fluxes in the partially and fully coupled experiments, their ocean circulation changes and the resulting circulation-driven ocean temperature components are also different (i.e., $v' \neq v'_p$ and $T_{\text{OO}} \neq T_{\text{OO}}^p$; subscript $p$ here denotes the partially coupled variables). Unlike $T_{\text{OO}}$, $T_{\text{OO}}^p$ does not also play a driving role in the air-sea interaction in the partially coupled experiment, since it is excluded from the surface flux calculation.

The partially coupled analogs to Equations 5 and 6 for the surface- and circulation-driven anomalous ocean temperature components can be also be written as

$$\frac{\partial T_{\text{OO}}}{{\partial t}} = Q'_{A} - v_p \cdot \nabla T_{\text{OO}}$$

$$\frac{\partial T_{\text{OO}}^p}{{\partial t}} = -v'_p \cdot \nabla T_0 - v_p \cdot \nabla T_{\text{OO}}$$

It might be argued that the circulation differences between the fully and partially coupled experiments could also have an impact on the atmosphere-driven responses simulated in the experiments, since the ocean circulation strength advecting their surface-driven temperature responses to $Q'_{A}$ are different (i.e., $T_{\text{OO}} = T_{\text{OO}}^p + T_{\text{OO}}$, in Equations 5 and 8). However, the impact of the circulation differences on the surface-driven components is higher order and small. Differencing Equations 5 and 8 which gives $(Q' - Q'_{A} - (v' - v'_p) \cdot \nabla (T_{\text{OO}} - T_{\text{OO}}^p))$, the second term is a higher-order term, which indicates that the surface-driven ocean temperature anomalies are more sensitive to surface flux difference than circulation change differences.

As a practical detail, we note that the partial coupling method described here is used only for the coupling between atmosphere and ocean models. As in the fully coupled experiment, the atmosphere-ocean-ice flux exchange in the partially coupled experiment is computed with the actual atmosphere and ocean temperatures. Using the actual ocean temperature that includes both surface- and circulation-driven components ensures that energy is conserved in the partially coupled experiment and the ocean temperature does not fall unrealistically below the freezing temperature. Since the ocean-ice heat flux exchange occurs at constant freezing and melting temperatures, it requires adjusting the ocean temperature for freezing and melting; adjusting a partial ocean temperature can cause unrealistically low ocean temperatures and an imbalance in the ocean heat budget. However, the ocean heat used for sea ice growth and melt could also be decomposed into surface- and circulation-driven components using the tracers in the fully and partially coupled experiments. This decomposition is necessary for partitioning the anomalous melt fluxes also into atmosphere- and ocean-driven components (discussed in section 3.2) and diagnosing the impact of the two-way air-sea interactions on the sea ice response. The partitioning of the ocean heat used for sea ice growth and melt and their impact on the sea ice response in the partially and fully coupled experiments is discussed in detail in a future study.

### 2.4. Ocean-Driven Anomalous Surface Flux Simulation and Validation

The partitioning of the air-sea interaction derived in Equations 1–3 assumes that the fully coupled response is approximately linear—equivalently that the atmosphere- and ocean-driven responses simulated separately will sum to the total response in the fully coupled experiment. It is important to verify this assumption and validate the atmosphere- and ocean-driven decomposition described above. To do this, we employ the previously used partial coupling approach of Bouttes et al. (2014) and Gregory et al. (2016) to simulate the ocean-driven response independently (referred to here as an “ocean-driven partially coupled experiment”) and compare that result with one derived by differencing the fully coupled and atmosphere-driven partially coupled experiments. The main difference between this partially coupled experiment and the one described
In section 2.3 is that the anomalous air-sea interactions are driven only by the circulation-driven or driving ocean temperature anomalies $T'_{Oo}$ rather than by an atmosphere surface temperature perturbation, so that the anomalous surface heat fluxes simulated are ocean-driven $Q'_O$, as defined in Equation 3 (Figure 1c).

In this experimental setup, no perturbation is applied to the atmosphere model, but additional anomalous surface fluxes $\tilde{Q}$ (defined in the next paragraph) are added to the surface fluxes felt by the ocean model (i.e., in addition to the normal surface flux exchange between the atmosphere and ocean, $Q_{cpl}$). Ocean circulation changes are thus produced by the anomalous surface fluxes added for the ocean model. As in the previous experiments, two tracers $P_1$ and $P_2$ are again constructed in the ocean model to isolate the surface-driven response to $\tilde{Q}$ (denoted as $T'_{Oo}$) and the circulation-driven ocean temperature anomalies produced by the circulation changes in the ocean model ($T'_{OO}$), respectively. Tracer $P_1$ is also used for computing the surface flux exchanged between the atmosphere and ocean (i.e., $T'_{cpl} = P_2$).

As in the previous experiments, at the initial time, $P_1$ and $P_2$ are set equal to zero and the baseline ocean temperature field ($T_O$), respectively. However, here, they are forced with the anomalous surface heat fluxes prescribed for the ocean model and the surface heat flux exchanged between the atmosphere and ocean, respectively (i.e., $Q_1 = \tilde{Q}$ and $Q_2 = Q_{cpl}$). Tracers $P_1$ and $P_2$ thus represent different but related quantities to the ones in the previous experiments (compare with previous tracers forcings in Table 1). At the initial time, $P_2 = T_O$ and $Q_{cpl} = \tilde{Q}$; however, as the anomalous surface fluxes added to the flux felt by the ocean ($Q_{cpl} + \tilde{Q}$) produce ocean circulation changes, $P_2$ will deviate from $T_O$ and also include a circulation-driven ocean temperature anomaly ($T'_{OO}$) over time. Since $P_2$ is also used to compute the surface flux exchange ($Q_{cpl}$), the circulation-driven anomaly in $P_2$ will also cause additional ocean-driven anomalous surface flux exchange, over time (i.e., $Q_{cpl} = \tilde{Q} + Q_O$). $P_2$ is also forced with $Q_{cpl}$, so that $P_2$ will also include additional surface-driven response to $Q'_O$, over time (i.e., $P_2 = T_O + T'_{OO} + T'_{Oo}$). The net anomalous ocean temperature participating in the air-sea interaction thus includes both driving and damping ocean temperature components ($T'_{OO}$ and $T'_{Oo}$) as defined in Equation 3. Note that $Q'_O$ also drives a damping temperature response in the atmosphere ($T'_{A0}$) as defined in (3). Since the net surface flux felt by the ocean model is $\tilde{Q} + Q_{cpl}$, the net ocean circulation change results from both prescribed and coupled anomalous fluxes $\tilde{Q}$ and $Q'_O$, and the tracers $P_1$ and $P_2$ sum up to the net ocean temperature in the experiment (i.e., $T_O = T_O + T'_{OO} + T'_{Oo} + T'_{Oo}$).

In order to reproduce the ocean-driven component of the fully coupled experiment, it is necessary to restrict the imposed anomalous fluxes for the ocean model $\tilde{Q}$, to the atmosphere-driven anomalous surface fluxes derived from the previous partially coupled simulation (i.e., $\tilde{Q} = Q'_A$), so that the total anomalous fluxes felt by the ocean in this partially coupled simulation is $Q' = Q'_A + Q'_O$. Choosing a different anomalous surface flux would lead to overestimating or underestimating the ocean circulation response and the resulting ocean-driven surface fluxes. For example, using anomalous surface fluxes from a fully coupled experiment would cause a double counting of the ocean-driven response. This ocean-driven partial coupling approach thus relies on appropriate isolation of the atmosphere’s role for an accurate isolation of the ocean-driven response, while the atmosphere-driven one can be used to isolate both responses, by simulating the atmosphere-driven component and then inferring the ocean-driven response as the residual between that simulation and the fully coupled simulation provided the fully coupled approach is linear. We demonstrate below that the linear assumption is valid, and this indicates that the approach suggested in section 2.3 provides a convenient and accurate decomposition of atmosphere and ocean responses to an external forcing.

**2.5. Summary of Model Configuration and Experiments**

We used the coupled community Earth System Model version1.2 (CESM 1.2). The standard-coupled CESM consists of the Community Atmospheric Model version 5 (CAM5, Neale et al., 2010), the Parallel Ocean Program version 2 (POP2, Danabasoglu et al., 2012), the Community Land Model version 4 (CLM4, Oleson et al., 2010), and the Community Ice CodE (CICE, Hunke et al., 2010). The horizontal resolution used for CAM5 and CLM4 is $1.5^\circ \times 0.9^\circ$, with the atmospheric component having 30 vertical levels. CICE and POP run on a nominally $1^\circ$ resolution with a displaced-pole grid (with the north pole singularity centered over Greenland); POP has 61 vertical levels.
An equilibrated 100-year period (years 1001–1100) of a control simulation with preindustrial CO$_2$ concentrations from the (CTRL; see Table 2) is used as the baseline simulation. Three perturbed simulations, initialized from year 1001 of the CTRL simulation are performed. The first is a 100-year long standard fully coupled simulation (FULL) that supports the complete air-sea interaction and all components of the ocean temperature field are used for coupling with the atmosphere but with quadrupled preindustrial CO$_2$ concentrations applied. The second perturbed simulation is an “atmosphere-driven” partially coupled simulation (PARTatm), that is also 100-year long and forced with quadrupled CO$_2$ concentration, but only the atmosphere-driven response to the CO$_2$ forcing is allowed as described in section 2.3. The ocean-driven response to CO$_2$ quadrupling is derived by differencing the FULL and PARTatm experiments. An additional perturbed “ocean-driven partially coupled” simulation (PARTocn) is also performed to verify the validity of ocean-driven response derived from FULL and PARTatm, as described in section 2.4. The PARTocn experiment is forced by prescribing all anomalous surface flux components from PARTatm for the ocean model. Since it serves only to verify ocean-driven response, PARTocn is run only for 50 years.

Two additional temperature-like tracers were introduced in each perturbed experiment to track the surface- and circulation-driven ocean temperature anomalies (see Table 1 and appendix B). Model output anomalies from the fully and partially coupled experiments are defined with respect to the monthly averages from the control experiment. The baseline surface fluxes used for forcing the tracers and ocean temperature added for coupling in the atmosphere-driven partially coupled experiment are similarly chosen to be monthly averages from the control experiment. Although other baseline choices are possible when the partial coupling method here is applied to study other questions (e.g., monthly climatology is used in Garuba, Lu, Singh, et al., 2018), choosing the interannually varying monthly averages from the control experiment allows for the isolation of anomalies due to CO$_2$ increase alone (this would be the baseline obtained if the CO$_2$ perturbation was not turned on in the perturbed experiments).

It is also necessary to replicate similar baseline surface fluxes in the PARTatm, PARTocn, and FULL experiments, so that only the anomalous surface fluxes differ between these experiments. In the FULL and PARTocn experiments, daily averages of the baseline and anomalous ocean temperature that are computed online (i.e., $T_0 = T_{0o} + P_2$ and $P_2 = T_{0o} + T_{0o}$) are passed to the coupler daily to compute the baseline and anomalous surface fluxes that are simulated in the experiments (i.e., $Q_{oc} = \overline{Q} + Q^0$ or $\overline{Q} + Q_{oc}$). This baseline ocean temperature therefore must include daily, seasonal, and interannual variations. In contrast, in the PARTatm experiment, the baseline ocean temperature used for computing the baseline surface fluxes is supplied as input. It is therefore necessary to use a time averaged baseline temperature that is as close as possible to the one simulated in FULL and PARTocn experiments. Ideally, daily averages of the surface temperature from the control experiment would be closest to the baseline ocean temperature that is coupled in the FULL experiment. However, due to the lack of daily averaged output from the control experiment, monthly averaged surface temperature fields from the concurrent month in the control experiment but interpolated to the daily coupling time in the PARTatm experiment are used. We evaluate the impact of this choice of baseline, using another short (10-year) partially coupled experiment with similar baseline ocean temperature component supplied as input but without CO$_2$ increase. A comparison of the surface heat fluxes in the original

| Table 2 | Summary of Experiments |
|---|---|---|---|---|
| Acronym | Name | Description | Length | $T_{oc}$ (ocean temperature coupled with the atmosphere) |
| CTRL | Control fully coupled | Preindustrial CO$_2$ | 100 years | $T_0$ |
| FULL | Perturbed fully coupled | 4xCO$_2$ forcing | 100 years | $T_0 + T_{0o} + Q_{oc}$ |
| PARTatm | Perturbed partially coupled (Atmosphere-driven) | 4xCO$_2$ forcing | 100 years | $T_0 + T_{0o}$ |
| PARTocn | Perturbed partially coupled (Ocean-driven) | Preindustrial CO$_2$ + prescribed atmosphere-driven anomalous surface fluxes (heat, freshwater, and momentum) | 50 years | $T_0 + T_{0o} + Q_{oc}$ |

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control and this partial coupled control experiments shows that the bias due to this baseline is indeed small about 3% of the CO₂ induced anomalies in the PARTatm experiment (0.21 W m⁻² compared to 7.40 W m⁻²; first 10-year average). The bias also occurs largely in the first year and quickly dissipates after the first year (Figures S1 and S2).

3. Results

3.1. Anomalous Temperature and Surface Heat Flux Responses

3.1.1. Baseline Heat Flux Pattern and Components

The baseline surface heat exchange and its components in the CTRL experiment are shown in Figure 2. The net surface heat flux pattern (panel a) includes surface heat gain in the equatorial regions, and heat loss in the western tropical Pacific and Indian Oceans and along the western boundary currents (Kuroshio and Gulf stream and Green land current) and around 60°S. This baseline heat flux pattern causes a negligible net global heat input of 0.06 W m⁻² into the ocean.

The baseline heat flux pattern is the net result of surface heat gain through the radiative component and heat loss through the latent and sensible components (compare Figures 2b–2d). Radiative heat gain is greatest in the tropics and decreases poleward (Figure 2b); latent heat loss is also generally greater in the tropics and decreases poleward, but the latent heat loss along the equator is weaker than the rest of the tropical regions (especially on the Eastern Pacific and Atlantic; Figure 2c). The relatively weak latent loss here results in the net surface heat gain in the equatorial regions. Weak latent heat loss along the equator results from the relatively cool ocean temperatures there. Equatorial ocean temperatures are cooler despite the net surface heat gain there, because of the wind-driven ocean upwelling caused the easterly winds along the equator. Outside the equatorial regions, latent heat loss is greatest on the western tropical regions, causing the net surface heat...
heat loss in these regions. Heat loss by the sensible component is generally weaker but contributes most strongly around the north Pacific and Atlantic western boundary currents (Kuroshio and Gulf stream), around the Bering sea and Greenland current, and around 60°S in the Southern Ocean (Figure 2d); this adds to the net heat loss by the ocean in these regions.

3.1.2. Atmosphere- and Ocean-Driven Anomalous Surface Heat Flux Components

On the other hand, the anomalous surface heat flux patterns caused by CO₂ quadrupling in the FULL and PARTatm experiments are different (Figures 3a and 3b). Recall that the anomalous surface heat fluxes in the FULL experiment include atmosphere- and ocean-driven components (i.e., \( Q' = Q'_{A} + Q'_{O} \)), while those in PARTatm are atmosphere-driven only (i.e., \( Q'_{A} \); see section 2.3). The atmosphere-driven surface heat flux pattern in the PARTatm experiment includes large heat gain in the Southern Ocean near 60°S, along the equator and along western boundaries currents in the high North Pacific and Atlantic (Kuroshio and Gulf streams) and Bering sea (Figure 3b). Except for the equatorial regions, these regions also have the large heat gain in the FULL experiment, especially in the Southern Ocean region, indicating that the fully coupled anomalous surface heat fluxes in these regions are largely atmosphere-driven (Armour et al., 2016; Fyfe et al., 2007; Garuba, Lu, Liu, et al., 2018; Marshall et al., 2015). However, the fully coupled anomalous surface heat flux response is different in the equatorial and subtropical regions, and in the subpolar Atlantic, hinting at an ocean-driven response in these regions.

The ocean-driven anomalous heat flux pattern is inferred by differencing the surface heat flux patterns in FULL and PARTatm (Figure 3c). Though with different patterns, the net global and time averaged heat flux pattern in the FULL and PARTatm experiments are similar (3.95 and 4.0 W m⁻², respectively), indicating that the ocean-driven anomalous surface heat flux contributes very little to the net ocean heat gain. The

![Figure 3](image-url)

Figure 3. Anomalous surface heat fluxes and its components, averaged over the first 50 years in the perturbed experiments: (a) Total \( Q = Q_{A} + Q_{O} \), from the FULL experiment, (b) atmosphere-driven component \( Q_{A} \), from the PARTatm experiment, (c) ocean-driven component \( Q_{O} \), inferred from the difference between the FULL and PARTatm experiments, and (d) ocean-driven component \( Q_{O} \), from the PARTocn experiment; unit = W m⁻². Positive sign indicates flux is into the ocean.
very small contribution of the ocean-driven component to the net heat uptake by the ocean can be explained by the redistributive property of the circulation-driven ocean temperature component driving it. This anomalous ocean temperature component is caused by the rearrangement of the baseline ocean temperature gradients by its circulation changes, therefore cooling in some regions must be accompanied by warming in other regions due to this redistribution (see section 2.2). The ocean-driven heat flux pattern driven by this anomalous surface temperature pattern thus includes heat losses and gains that nearly cancel out each other.

The ocean-driven anomalous heat flux pattern includes large heat loss in the equatorial regions and heat gain in the subpolar Atlantic and relatively weaker heat gain in the subtropics (Figure 3c). This heat flux pattern is consistent with what is shown in earlier studies (Garuba & Klinger, 2016; Garuba, Lu, Liu, et al., 2018; Gregory et al., 2016; Winton et al., 2013) and also largely reproduced by the independent simulation of the ocean-driven response in the PARTocn experiment (compare Figures 3c and 3d). The similarity in the derived and simulated ocean-driven patterns verifies our conjecture that the fully coupled anomalous surface heat flux response is indeed linear and can be decomposed into atmosphere- and ocean-driven responses using our partial coupling approach employed in the PARTatm experiment. Since the decomposition appears linear and representative, it is useful to further decompose the atmosphere-driven anomalous surface heat fluxes (PARTatm) and the derived ocean-driven ones (FULL-PARTatm) into contributions from the sensible, latent, and radiative fluxes (Figure 4).

The radiative and sensible components of the atmosphere-driven anomalous surface heat fluxes cause heat gain, while the latent component causes heat loss in most regions (Figures 4a, 4c, and 4e). The exception to

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**Figure 4.** Anomalous radiative, latent, and sensible surface heat flux components averaged over the first 50 years: Atmosphere-driven components from the PARTatm experiment (a, c, and e), and the ocean-driven components inferred from the difference between the FULL and PARTatm experiments (b, d, and f), respectively; unit = W m$^{-2}$. Positive sign indicates flux is into the ocean.
these are in the equatorial and polar regions, with the anomalous latent heat gain and sensible heat loss, respectively. Except in the equatorial regions, the atmosphere-driven anomalous radiative heat gain largely cancels out the latent heat loss (compare Figures 4a and 4c), so that the net atmosphere-driven heat gain pattern is largely contributed by the sensible component (compare Figures 3b and 4e). Identifiable contributions of the sensible component to the net atmosphere-driven heat gain pattern can be seen in the northern Pacific and Atlantic western boundary currents (Kuroshio, Gulf stream), around Bering sea, Green land current, and around 60°S in the Southern Ocean; elsewhere, atmosphere-driven sensible heat gain is weak. In the equatorial regions, the net atmosphere-driven anomalous heat gain is contributed by both latent and radiative components (compare Figures 3b, 4a, and 4c), while over the polar regions, new open waters due to the sea ice loss in the PARTatm experiment cause anomalous radiative heat gain and latent and sensible heat losses.

The greater contribution of the sensible component to the net atmosphere-driven anomalous heat flux pattern is consistent with the atmospheric origin of this pattern. According to Equation 1, baseline sensible and latent heat losses by the ocean in the CTRL experiment (Figures 2c and 2d) indicate warmer baseline ocean surface temperatures than atmosphere surface temperatures on average (i.e., \( T'_O > T'_A \)). With CO2 increases, net (i.e., baseline plus anomaly) radiative surface heat gain increase, but sensible heat loss decrease in the PARTatm experiment, except in the polar regions (compare baselines in Figures 2b and 2d with atmosphere-driven anomalies in Figures 4a and 4e). Even though both atmosphere and ocean surface temperatures participating in the air-sea interaction increase in the PARTatm experiment, the decrease in the
net sensible heat loss in this experiment indicates that the change in the air-sea temperature difference in this experiment is driven by an atmosphere (rather than an ocean) surface temperature increase. An increase in ocean surface temperature would increase (rather than reduce) the underlying baseline air-sea temperature difference and sensible heat loss. The ocean surface temperature increase is instead a response to the atmosphere-driven anomalous radiative and sensible heat gains, which plays a damping role in the air-sea interaction according to Equation 3 by increasing the net latent heat loss almost everywhere (compare Figures 2c and 4c). However, in the equatorial regions, latent heat loss decreases despite the ocean surface temperature increase there (compare Figures 4c, 5c, and 5d), but this decrease in the equatorial latent heat loss also originates in the atmosphere. Similar to the baseline latent heat loss behavior here, this decrease can be explained by the tight coupling between SSTs and surface winds in this region. As SSTs increase, the zonal SST gradient along the equator also weakens and thereby causes a weakening of equatorial easterlies also; weaker winds reduce evaporation rates and latent heat loss along the equator (Vecchi et al., 2006; Xie et al., 2010).

On the other hand, the ocean-driven anomalous heat flux pattern is largely contributed by the latent heat component (compare Figures 3c and 4d). The ocean-driven anomalous heat gain in the subpolar Atlantic and subtropical regions and heat loss over the equatorial regions are primarily from the latent heat component. There are also notable contributions from the sensible component to the ocean-driven heat flux in the subpolar Atlantic and from the radiative component in the equatorial region, but their contributions are generally weaker (Figures 4b and 4f). Notably, there is an ocean-driven radiative heat loss along the equator that is flanked by radiative heat gains north and south of the heat loss. This radiative pattern is explained by an ocean-driven precipitation change (Figure 10b; see section 3.2 for further discussion). Overall, the greater contribution of the latent heat component to the ocean-driven heat flux pattern is consistent with its oceanic origin. Latent heat fluxes more sensitive to ocean temperature changes than any other surface heat flux component; consistently ocean-driven anomalous latent and sensible heat flux patterns are largely explained with the ocean-driven anomalous surface temperature pattern.
This result is consistent with earlier studies linking the pattern of evaporation changes in global warming experiments to SST pattern in the extratropical regions (Long & Xie, 2015). Our analysis here shows that the SST pattern producing the evaporative changes are driven by ocean circulation changes rather than atmosphere-driven surface heat fluxes.

3.1.3. Atmosphere and Ocean-Driven Temperature Responses to CO₂ Increases

Similar to the anomalous surface heat fluxes, the ocean temperature response in the FULL experiment can also be partitioned into atmosphere- and ocean-driven components (Figure 5). Following the definitions in Equation 3, the atmosphere- and ocean-driven components are the net ocean temperature changes participating in the atmosphere- and ocean-driven air-sea interactions, respectively (i.e., \( T'_{OQO} \) and \( (T'_{OQO} + T'_{OO}) \)), and these can provide insight into the atmosphere- and ocean-driven anomalous surface heat fluxes. Note that the atmosphere- and ocean-driven partitioning of the anomalous ocean temperature is different from its surface-and circulation-driven partitioning (Figure 6). The surface-driven component in the FULL experiment, for example, includes ocean temperature responses to both atmosphere- and ocean-driven surface heat flux anomalies (i.e., \( T'_{OQO} + T'_{OO} \); Figure 6b) and so cannot be the atmosphere-driven component. Compared to the one in PARTatm, this surface-driven temperature pattern includes a greater warming response in the subpolar Atlantic and cooler temperature response in the tropics due to the additional ocean-driven surface heat gain and heat loss over in these regions in the FULL experiment (compare Figures 6a, 6b, 3a, and 3b). Similarly, the circulation-driven ocean temperature component in the FULL experiment (\( T'_{OO} \)) is not equivalent to the ocean-driven component because it excludes the damping ocean temperature response to ocean-driven anomalous surface heat fluxes (\( T'_{OQO} \)) and so cannot fully explain the ocean-driven anomalous surface fluxes.
On the other hand, the surface-driven ocean temperature response in the PARTatm experiment is equivalent to the atmosphere-driven ocean temperature component, since it is the only ocean temperature component participating in the atmosphere-driven air-sea interaction (i.e., $T'_{\text{OQA}}$ in Equation 3; compare Figures 5c, 5d, and 6a). Recall that this ocean temperature response component acts only to damp the atmosphere-driven anomalous surface fluxes (section 2.1). However, the atmosphere-driven ocean temperature response pattern is distinct from the atmosphere-driven surface heat gain pattern that drives it (compare Figures 3b, 5c, and 5d). Though the atmosphere-driven anomalous surface heat fluxes have similar magnitudes in both hemispheres, the atmosphere-driven warming response is greater in the northern hemisphere than the southern hemisphere, with even some cooling in the Antarctic. Atmosphere-driven ocean temperature responses along the equator and Southern Ocean are also weak, despite large atmosphere-driven surface heat gain in these regions. The differences between the atmosphere-driven surface heat flux and ocean temperature response patterns are explained by the ocean heat convergence term in Equation 8 ($v_p \cdot \nabla T'_{\text{OQA}}$), which determines the damping rates of the atmosphere-driven surface heat fluxes, such that regions of large heat uptake are not necessarily regions of large warming (Armour et al., 2016; Marshall et al., 2015). Nevertheless, this atmosphere-driven pattern is very similar to the ocean temperature response pattern obtained from slab ocean experiments or fixed ocean current experiments (e.g., Figures S5 and 6 in Garuba, Lu, Liu, et al., 2018; Winton et al., 2013), adding confidence to the validity of this partitioning. Note though that the magnitude of the atmosphere-driven response in the PARTatm experiment here is weaker than the one in a slab experiment because of the anomalous ocean heat convergence response that is included in the PARTatm experiment (i.e., additional $v_p \cdot \nabla T'_{\text{OQA}}$ in PARTatm compared to $v \cdot \nabla T$ in a slab).

In contrast, the ocean-driven temperature component is the difference between the total ocean temperature response in the FULL experiment and its atmosphere-driven component (i.e., $T'_{\text{OO}} + T'_{\text{OQA}}$ in Equation 3;...
Figures 5e and 5f). Recall that this ocean-driven temperature pattern is the net ocean temperature anomaly that is coupled with the atmosphere, and over time it includes both circulation-driven (or driving) anomalous ocean temperature and surface-driven (or damping) ocean temperature responses to the ocean-driven surface heat fluxes. Unlike the atmosphere-driven one, the ocean-driven temperature pattern is generally cooler in the northern hemisphere than the southern hemisphere; with the largest cooling occurring over the subpolar Atlantic and an El-Niño-like warming over the equatorial and general warming in the Southern Ocean regions (Figures 5e and 5f). This temperature pattern is similar to the surface temperature response to AMOC shutdown in the studies of Vellinga and Wood (2002) and Zhang and Delworth (2005) and is shown to cause the time evolution of the net surface temperature response and ocean heat uptake patterns and climate sensitivity of the fully coupled response (Andrews et al., 2015; Garuba, Lu, Liu, et al., 2018; Rugenstein et al., 2016). The large subpolar Atlantic cooling in this ocean-driven pattern drives the large ocean-driven anomalous heat gain in the subpolar Atlantic which compensates the cooling. Similarly, the equatorial warming in this pattern drives the ocean-driven anomalous heat loss over the tropics (compare Figures 3c, 5e, and 5f).

The underlying ocean temperature anomaly pattern that produces the net ocean-driven ocean temperature response is the circulation-driven ocean temperature anomaly that is excluded from the air-sea interaction in the PARTatm experiment (i.e., \( T_{O0} \); Figure 6c). When this circulation-driven temperature pattern is coupled with the atmosphere in the FULL or PARTocn experiments, it drives the ocean-driven anomalous surface heat fluxes \( (Q'_O) \) and further ocean circulation changes, and the surface- and circulation-driven ocean temperature responses associated with them (i.e., \( T_{O0} \) and \( (T'_{O0} - T_{O0}) \)), respectively, which are included in the net ocean-driven temperature response. The ocean-driven temperature fluxes act to compensate the uncoupled circulation-driven ocean surface temperature anomalies through the surface-driven temperature response to \( Q'_O \). The compensating impact of the ocean-driven surface heat loss in the tropics (Figure 3c), thus explains a relatively weak tropical warming response in the ocean-driven temperature pattern compared to the large tropical warming in the uncoupled circulation-driven temperature pattern (compare Figures 6c, 5e, and 5f). However, similar circulation-driven tropical warming responses are seen in the PARTatm and FULL experiments, indicating that \( Q'_O \) does not cause additional circulation changes in the tropical region (Figures 6c and 6d).

In contrast, the large ocean-driven anomalous heat gain in the subpolar Atlantic (Figure 3c) does not appear to compensate the circulation-driven cooling in this region, and the ocean-driven pattern has even greater cooling in the subpolar Atlantic compared to that of the uncoupled circulation-driven temperature pattern (compare Figures 5e, 5f, and 6c). This further cooling of the subpolar Atlantic is explained by additional circulation changes caused by the ocean-driven anomalous heat gain in the subpolar Atlantic. The weakening of the Atlantic Meridional Overturning Circulation (AMOC) is known to be very sensitive to the anomalous surface heat fluxes in the subpolar Atlantic and also associated with ocean surface temperature cooling in this region (Garuba & Klinger, 2016; Garuba, Lu, Liu, et al., 2018; Gregory et al., 2016; Winton et al., 2013). As a result, the ocean-driven surface heat gain in the subpolar Atlantic causes further weakening of the AMOC, which in turn causes further circulation-driven cooling of the region in the FULL experiment (Figures 6c and 6d; see section 3.3 for further discussion of circulation changes). The additional circulation-driven cooling apparently exceeds the compensating warming effect of the ocean-driven surface heat gain in this region (Figures 6b and 6d), so that the resultant ocean-driven temperature response is even cooler than the uncoupled circulation-driven cooling in this region. This net cooling response explains why the ocean-driven surface heat gain is greater here than anywhere else. This coupled positive feedback on the AMOC weakening is also noted in the study of Gregory et al. (2016).

### 3.2. Anomalous Surface Salinity and Momentum Flux Responses

Anomalous surface salinity and momentum fluxes in the FULL experiment can also be partitioned into atmosphere and ocean-driven components. Similar to the anomalous surface heat fluxes, the salinity fluxes in PARTatm are atmosphere-driven by design. However, partitioning the salinity fluxes is slightly different because of sea ice changes. The melt component of the salinity fluxes can include an ocean-driven component, since sea ice melt flux anomalies are caused in part by ocean temperature changes. The anomalous melt fluxes in PARTatm thus include a component that is caused by the uncoupled circulation-driven anomalous ocean temperature which is excluded from atmosphere-driven component by definition.
Similar to the ocean temperature partitioning, the atmosphere-driven melt flux component must include only the surface-driven anomalous melt flux component in the PARTatm experiment; this component would be due to the sea ice melt or growth changes caused only by atmosphere-driven surface heat input into the ocean (i.e., caused by $T_{00}$). Similarly, the ocean-driven melt flux component is derived from the difference between the total melt flux anomalies in the FULL experiment and the atmosphere-driven component; this component would be due to sea ice melt and growth changes caused by the total ocean circulation change in the FULL experiment and the ocean-driven anomalous surface heat input into the ocean (i.e., caused by $T_{00} + T_{00}$). To partition the melt fluxes, the anomalous ocean heat used for sea ice melt and growth is first partitioned into surface- and circulation-driven components using the tracers in the experiments (details are discussed in another study). The melt flux components are obtained offline as the weighted contribution of the ocean-ice heat flux components to the net melt fluxes.

Unlike anomalous surface heat fluxes, the net atmosphere-driven and total anomalous surface salinity fluxes in the PARTatm and FULL experiments are similar (Figures 7a and 7b). This similarity indicates that the surface fresh water flux changes in the FULL experiment are largely atmosphere-driven, though there are also notable contributions from the ocean-driven component (Figures 7c and 7d). The atmosphere-driven salinity flux pattern includes a freshening along the equatorial, Southern Ocean and Northern polar regions, and salinification in the tropical/subtropical regions, especially in the Atlantic (Figure 7b). While the ocean-driven pattern includes a freshening along the equatorial and subpolar Atlantic regions, and salty anomalies north of the equatorial freshening band (with weaker salty anomalies south of it; Figure 7c). The polar regions also show an interesting dipole salinity flux anomaly, with freshening around the poles and salty anomalies equatorward of the polar freshening band (Figure 7c). This ocean-driven pattern is generally reproduced in the PARTocn experiment (Figure 7d), indicating that fully coupled surface salinity flux response is also linear.

Consistent with its atmospheric origin, the atmosphere-driven salinity flux pattern is largely from the precipitative component, rather than the evaporative or sea ice melt components (compare Figures 8a, 8c, and 8e). The atmosphere-driven freshening along the equator and Southern Ocean and salinification of the subtropics are caused by increases and decreases in precipitation, respectively (Figure 8a). Some freshening along the equator is also caused by a decrease in evaporation, though much smaller (Figure 8c, see also discussion in section 3.1.2). Increases in precipitation in the tropics and its decreases in the subtropics is consistent with the known wet-get-wetter precipitation response to global warming (Chou & Neelin, 2004; Chou et al., 2009; Held & Soden, 2006; Xie et al., 2010). The precipitation (rainfall and snow) increase in the Southern Ocean is also a known atmosphere-driven response to global warming that is associated with the poleward intensification of Southern Ocean winds (Frieler et al., 2015; Palerme et al., 2017; Uotila et al., 2007; Yin, 2005). This poleward shift of the Southern Ocean winds also explains the increase in melt around Antarctica and poleward increase in snow-ice growth (Figure 8e).

The ocean-driven anomalous salinity flux pattern is also largely caused by anomalous precipitation rather than evaporation (or latent heat loss by the ocean) (Figures 8b and 8d). However, since this anomalous precipitative pattern is ocean-driven, it might be explained by a southward shift of the ITCZ that would result from the large northern hemisphere cooling in the ocean-driven SST warming pattern (compare Figures 5e and 5f; see also section 3.3 for further discussion). Since the baseline position of the ITCZ is north of the equator, a southward shift of the ITCZ’s precipitation band causes an increase in precipitation (freshening anomalies) along the equator, and a decrease in precipitation (saltier anomalies) north of the equator.

This precipitative pattern is also consistent with the ocean-driven radiative surface heat flux response that would accompany the ITCZ shift (see section 3.1.2; compare Figure 4), with cloud radiative cooling along the equator and warming north of the equator. This ocean-driven precipitative pattern enhances the atmosphere-driven wet-get-wetter pattern in the FULL experiment, especially along the equator and northern subtropics (compare Figures 7a and 7b). As shown in the study of He and Soden (2017), the wet-get-wet precipitation response is also partly explained by the SST pattern. This SST-pattern-driven component might be equivalent to the ocean-driven precipitative pattern seen here; as discussed in the previous section, the net SST pattern comes largely from the ocean-driven component, especially over the tropics (compare Figures 5b, 5d, and 5f). In the subpolar Atlantic, the ocean-driven freshening is caused by a decrease in evaporation due to the ocean-driven cooling there (compare Figures 8d, 4d, 5e, and 5f). While in the polar
regions, the ocean-driven salinity dipole pattern is caused by the circulation-driven increase in ocean heat transport into the polar regions (Figure 8f).

Like the anomalous salinity flux response, the momentum flux anomalies are also largely atmosphere-driven (compare Figures 9, 10a, and 10b). Notable features of the atmosphere-driven wind stress response include dipole zonal wind anomalies and negative meridional wind stress around 45°S (indicating a poleward shift or intensification of the Southern Ocean westerlies) and positive zonal wind stress anomalies along the equator (indicating a weakening of the tropical easterlies). The poleward intensification of the Southern Ocean westerlies is a robust feature of global warming experiments and projections which is also shown to be caused by the direct radiative effect of CO₂ in agreement with the atmosphere-driven response (Deser & Phillips, 2009; Fyfe & Saenko, 2006; Fyfe et al., 2007; Menzel & Merlis, 2019; Saenko et al., 2005). The weakening of the equatorial easterlies also cause the wind-driven decrease in latent heat loss along the equator (Figure 4c; see also discussion in section 3.1.2).

The ocean-driven features include negative zonal wind anomalies in the subpolar Atlantic, which is likely caused by the cooling of the subpolar Atlantic. Negative and positive meridional wind anomalies north and south of the equator, respectively (indicating an ocean-driven southward shift of the ITCZ) (Figures 9, 10c, and 10d). This ocean-driven wind stress response is also largely reproduced in the PARTocn, thus indicating that the wind response is also linear.

3.3. Ocean and Atmosphere Circulation and Heat Transport Responses

The ocean circulation changes resulting from the atmosphere and ocean-driven anomalous surface flux components are examined using the meridional overturning circulation (MOC) anomalies (Figure 11).
These ocean circulation changes explain the circulation-driven ocean temperature changes in PARTatm and FULL experiments described in section 3.1.3 (Figures 6c and 6d). Atmosphere-driven poleward intensification of the Southern Ocean westerlies and weakening of the tropical easterlies (Figures 9a and 9b) produce similar ocean circulation response in the southern hemisphere and subtropical cells in the PARTatm and FULL experiments. The southern hemisphere circulation change pattern includes dipole anomalies around 45°S (positive and negative, south and north of 45°S, respectively), and a weakening of the subtropical cells (Figure 11). These similar ocean circulation responses in the Southern Ocean and tropics explain the similarity in the circulation-driven responses in these regions in both experiment (compare Figures 6c and 6d).

In contrast, the circulation change in the northern hemisphere is different in the two experiments. The northern hemisphere overturning circulation anomaly is due differences in the AMOC weakening. The AMOC response stabilizes in the PARTatm experiment but continues to weaken in the FULL experiment after twenty years (Figure 12). The AMOC response is also largely reproduced in the PARTocn experiment, indicating that this response is also linear (Figure 12). The AMOC weakening differences cause the large circulation-driven cooling in the subpolar Atlantic in the FULL experiment (Figures 6c and 6d), and the overall cooler northern hemisphere in the ocean-driven temperature pattern (Figures 5e and 5f). The circulation response differences suggest that coupling provides a positive feedback on the AMOC weakening response (Gregory et al., 2016). The mechanism of this coupled AMOC response is beyond the scope of this paper and will be explored in a future study.

AMOC weakening differences in PARTatm and FULL experiment also have important consequences on both ocean and atmosphere heat transports (Figure 13, blue and red lines). Similar to its AMOC, the ocean northward heat transport continues to weaken in the FULL experiment, while in PARTatm the weakening occurs largely in the first 20 years (compare blue lines in Figure 13). The atmosphere compensates for this

Figure 10. Same as Figure 9 but for meridional wind stress anomalies.
anomalous ocean heat transport by increasing its northward heat transport. As a result, the atmospheric northward heat transport in the FULL experiment increases more than that of the PARTatm experiment (compare red lines in middle and bottom panels of Figure 13). That the anomalous atmospheric northward heat transport is weaker in PARTatm than the FULL experiment, even though PARTatm has greater Arctic amplified warming (compare Figures 5a–5d), indicates that Arctic amplification is not a consequence of the increase in the atmospheric northward heat transport.

Another impact of the AMOC weakening response appears in the position of the ITCZ. As shown by the meridional wind anomalies, there is an ocean-driven shift in the position of the ITCZ (Figures 10c and 10d). The change in the position of the ITCZ can also be inferred from the sign of the anomalous cross equatorial heat transport. The anomalous atmosphere heat transport at the equator is southward (negative) in PARTatm, but northward (positive) in the FULL experiment, which implies a northward shift of the ITCZ in PARTatm and southward shift in the FULL experiment (Figure 13; red lines in bottom two panels). The southward shift of the ITCZ in the FULL experiment is consistent
with known impacts of AMOC weakening shown in the studies of Broccoli et al. (2006), Vellinga and Wood (2002), and Zhang and Delworth (2005).

The difference in the ITCZ response in PARTatm and FULL experiments can also be explained by the inter-hemispheric asymmetry of the atmosphere-driven and ocean-driven temperature response patterns in these simulations (recall Figures 5c–5f). The ITCZ have been shown to shift toward the hemisphere in which an extratropical heat source is applied (Kang et al., 2008, 2009). Since the atmosphere sees only the atmosphere-driven temperature pattern with greater warming in the northern hemisphere than the southern hemisphere in the PARTatm experiment (Figures 5c and 5d). The greater northern hemisphere warming in this pattern therefore explains the northward shift of the ITCZ in the PARTatm experiment. While in the FULL experiment, the atmosphere sees an additional ocean-driven ocean temperature pattern with a large northern hemisphere cooling and southern hemisphere warming (Figures 5e and 5f). This ocean-driven temperature anomaly pattern thus explains the southward shift of the ITCZ in the FULL experiment.

4. Summary and Discussion

This study presents the formulation, application, and validation of a partial coupling method that can be used to disentangle and isolate the atmosphere- and ocean-driven components of air-sea interactions in a
coupled climate model. The atmosphere-driven component is identified using a partially coupled framework where the ocean's impact is disabled by isolating and removing the anomalous ocean temperature that caused by ocean circulation changes from the air-sea interaction. Temperature-like tracers are introduced into the simulation to track the anomalous ocean temperature components that are driven by anomalous surface heat fluxes and ocean circulation changes, respectively. The tracers are also used in the calculation of the surface heat exchange in order to simulate the atmosphere-driven response. The ocean-driven response can then be inferred by comparing the response in the partially coupled simulation with that of a standard fully coupled one. This atmosphere- and ocean-driven response decomposition is verified with a third independent simulation designed to isolate the ocean-driven response (using a complimentary partial coupling approach, where ocean circulation changes caused by atmosphere-driven surface fluxes drives the air-sea interaction).

As an example of how the method can be used, we applied it to better understand the fully coupled model response to CO₂ quadrupling by partitioning the response into atmosphere- and ocean-driven components. Our analysis indicates that atmosphere-driven surface heat flux changes induced by CO₂ increases occur preferentially in the western boundary currents in the North Pacific and Atlantic (Kuroshio and Gulf stream), around Bering sea and Greenland and around 60°S in the Southern Ocean, and along the equatorial regions. This atmosphere-driven heat gain by the ocean occurs largely through a decrease in sensible heat loss by the ocean in the Southern Ocean, Kuroshio and Gulf stream and Green land and Bering sea regions, and a decrease latent heat loss along the equator. The ocean temperature response to this atmosphere-driven heat gain is an Arctic amplified warming pattern with overall greater warming in the northern hemisphere and some cooling in the Southern Ocean, which also causes a northward shift of the ITCZ. Other atmosphere-driven changes include: weakening of the easterlies along the equator, a poleward intensification of the Southern Ocean westerlies, which also cause an increase in precipitation in the equatorial and Southern Ocean regions, and decrease in the subtropical precipitation.

Ocean-driven anomalous surface heat fluxes, on the other hand, occur largely through the latent heat flux changes and include large heat gains in the subpolar Atlantic, weaker heat gains in the subtropical oceans and heat loss along the equator. This anomalous heat flux pattern is driven by the interaction of the atmosphere with a circulation-driven ocean temperature anomaly pattern, that results from the AMOC weakening in the Atlantic, and the wind-driven subtropical circulation changes in the Pacific. In contrast to the atmosphere-driven temperature response pattern, the net ocean-driven temperature anomaly pattern which drives this interaction includes a large cooling of the northern hemisphere, especially in the subpolar Atlantic, and an El Niño-like warming along the equator, and an overall warming of the southern hemisphere. The large inter-hemispheric asymmetry in this ocean-driven temperature response pattern, also causes a southward shift of the ITCZ and the precipitation band associated with it, and large increases in the atmosphere's northward heat transport which acts to compensate the decrease in the ocean northward heat transport caused by the weakening of the AMOC.

Some atmosphere- and ocean-driven responses identified here using the partial coupling method are consistent with earlier studies that use stand-alone frameworks, increasing confidence in the validity and robustness of the decomposition. Large atmosphere-driven heat gain in the Southern Ocean and the poleward shift of the winds are shown to be robust features of the global warming response, consistent with earlier studies (Armour et al., 2016; Gregory et al., 2016; Saenko et al., 2005; Yang et al., 2016). Atmosphere-driven increases in equatorially precipitation and decreases in subtropical precipitation are consistent with the robust wet-get-wetter precipitation response to global warming (Chou & Neelin, 2004; Chou et al., 2009; He & Soden, 2017; Held & Soden, 2006; Lu et al., 2007, 2008). The Arctic amplified surface warming response to the atmosphere-driven heat gain, despite producing a weak increase in atmospheric northward heat transport in the partially coupled experiment, is also consistent with a more important role of regional feedbacks than changes in northward heat transports in Arctic Amplification shown in the studies of Hwang et al. (2011); Kay et al. (2012); Stuecker et al. (2018). The large ocean-driven cooling of the northern hemisphere, especially over the subpolar Atlantic and El Niño-like warming in the tropical pacific and associated southward shift of the ITCZ, are consistent with known impacts of AMOC weakening and damping effects of the ocean on the ITCZ shifts (Broccoli et al., 2006; Collins et al., 2005; Garuba, Lu, Liu, et al., 2018; Green & Marshall, 2017; Gregory et al., 2016; Vellinga & Wood, 2002; Winton et al., 2013; Zhang & Delworth, 2005).
Due to its unique online simulation of both atmosphere- and ocean-driven responses, the partial coupling method also provides new insights into the fully coupled system. It provides information on the time evolution or timescales associated with the atmosphere- and ocean-driven responses, which cannot be deduced from stand-alone simulations. For example, the AMOC response in the partially and fully coupled experiments suggests that the AMOC weakening response to CO$_2$ induced anomalous surface heat fluxes alone, stabilizes after about twenty years, but the atmospheric coupling with this response provides a positive feedback that destabilizes the AMOC weakening even further; this coupled response also has a stabilizing effect on sea ice loss after about twenty years. The mechanism of the coupled AMOC and sea ice responses are beyond the scope of this study and will be explored elsewhere. The time evolution of the anomalous surface heat flux and temperature patterns and climate sensitivity were also shown to be ocean-driven responses in Garuba, Lu, Liu, et al. (2018), which used the partial coupling method described in much more detail here. Possible southward shifts of the ITCZ have been proposed using stand-alone simulations (Garuba et al., 2018), and here we show a continual southward shift of ITCZ position with time is indeed largely caused by the ocean using our decomposition.

Simulating surface flux and circulation changes online instead of prescribing boundary conditions or ocean heat fluxes and mixed layer properties in the partially coupled simulation also avoids possible double counting of the roles of the atmosphere and ocean and reduces the potential for flawed or inaccurate attribution. For example, although slab ocean model simulations are also designed to remove the impact of the ocean on sea interaction, they are constrained by prescribing ocean heat convergences usually derived from an unperturbed control simulation. As a result, they do not account for ocean heat convergence anomalies associated with ocean circulation and mixed layer changes (i.e., $v_p \cdot V^\circ_{OQA}$ term in Equation 8), which can be significant or even dominant in certain regions such as the North Atlantic and Arctic (Barsugli & Battisti, 1998). Slab models can therefore produce unrealistic surface flux anomalies and an overestimation of the upper ocean heat content change. This lack of realistic damping of upper ocean heat content by ocean circulation variations has lead to an attribution of the Atlantic multidecadal variability to atmospheric noise using slab simulations, but using the partial coupling approach here, we show that the multidecadal variability is most likely driven by ocean circulation changes in fully coupled simulations (Clement et al., 2015; Garuba, Lu, Singh, et al., 2018; Zhang, 2017). Avoiding the double counting of the roles of the atmosphere and ocean, our decomposition indicates that the fully coupled responses (surface flux components and ocean circulation responses) are indeed linear.

By incorporating evolving atmosphere and ocean components, the partial coupling described here also provides improved attribution compared to ocean-only studies that similarly remove surface flux anomalies produced by ocean circulation changes, using a surface boundary restoration in an ocean model (Garuba & Klinger, 2016, 2018). For example, El Niño-like warming in the tropical Pacific to surface heat flux changes, and the large cooling of the North Atlantic due to AMOC weakening were attributed to freshwater flux changes in the ocean-only study. However, this study provides a different interpretation; that the tropical warming response is wind-driven but the large AMOC weakening is largely due an ocean-driven surface heat flux response. The large freshening of the subpolar gyre is circulation-driven rather than surface-driven (not shown). The differing attribution arises from the treatment of the boundary forcing in the ocean-only simulation. The anomalous SST pattern used to drive the ocean-only model in Garuba and Klinger (2016, 2018) was derived from the SSTs changes in a fully coupled simulation that included the circulation-driven SST responses caused by surface wind and surface heat fluxes changes.

Some caveats and consequences of using the partial coupling strategy are worth mentioning. The computational costs of the partial coupling strategy is higher than studies using stand-alone simulations. The decomposition also depends on the definition of the base state which must be chosen carefully. Here, we used monthly averages from the control, interpolated to daily values in the partially coupled simulation, which removes daily variability of the fully coupled baseline. For the long time scales considered in this study, this is a good approximation, however for smaller time and spatial scales, this may not be a good approximation and daily averages of the baseline would be a better choice. Nevertheless, our results indicate that the partially coupling method here can be useful in teasing out the role of the atmosphere and ocean in climate variability on long timescales using a carefully chosen baseline. The partial coupling approach will also be useful for: 1) understanding the mechanisms of climate variability or climate response, or 2) assessing the
source of biases in fully coupled simulations since it can isolate the atmospheric, oceanic or coupled origin of such biases.

**Appendix A: Anomalous Ocean Temperature Tendency Equation**

The anomalous ocean temperature tendency Equation 4 is derived from the ocean temperature tendency equations for the perturbed and unperturbed or baseline states. The perturbed ocean temperature tendency equation may be written in terms of the baseline and anomalous components as

$$\frac{\partial (T_O + T'_O)}{\partial t} = Q + Q' - (v + v') \cdot \nabla (T_O + T'_O)$$  \hspace{1cm} (A1)

Similarly the baseline equation is

$$\frac{\partial T_O}{\partial t} = Q - v \cdot \nabla T_O$$  \hspace{1cm} (A2)

Expanding Equation A1 and subtracting Equation A2 gives

$$\frac{\partial T'_O}{\partial t} = Q' - v' \cdot \nabla T_O - v \cdot \nabla T'_O$$  \hspace{1cm} (A3)

**Appendix B: Tracer Equations**

The passive tracer decomposition method is similar to the ones used in several previous studies Banks and Gregory (2006); Xie and Vallis (2012); Bouttes et al. (2014); Marshall et al. (2015); Garuba and Klinger (2016); Gregory et al. (2016); Garuba, Lu, Liu, et al. (2018), and Garuba and Klinger (2018). This appendix provides more detail about the decomposition than the main description allows.

Tracers are used to isolate surface- and circulation-driven anomalous ocean temperatures component in the perturbed fully-coupled and partially-coupled simulations. The general equation for a temperature-like tracer $P_i$, with a surface source $Q_i$ that is advected by the ocean velocity field $v$, is given as

$$\frac{\partial P_i}{\partial t} = Q_i - v \cdot \nabla P_i$$  \hspace{1cm} (B1)

Two tracers formulations obeying Equation B1 are used to isolate surface- and circulation-driven components of the ocean temperature anomalies in each of the three simulations discussed in the manuscript. The choice of initialization and forcing terms and the role of the tracers differ according to the design of each experiments, as summarized in Tables 1 and 2. The rationale for the underlying basic formulation and the choices for specific experiments are described in more detail in the following paragraphs.

The first tracer denoted as $P_1$ tracks the temperature anomaly driven by the surface heat flux perturbation in the experiments. The surface source $Q_1$ for $P_1$ is therefore chosen to be a surface heat flux anomaly which is either computed (as in the FULL or PARTatm simulations) or prescribed (as in PARTocn). The surface heat flux anomaly is computed online by subtracting a baseline surface heat flux derived from the unperturbed control simulation, from the coupled surface heat fluxes in the perturbed simulations (i.e., $Q_{\text{cl}} - Q$), such that for the FULL and PARTatm simulations, $Q_1$ for tracer $P_1$ is $Q'$ and $Q'_{\text{A}}$, respectively. Since $P_1$ tracks only the anomalous ocean temperature produced surface heat flux anomalies starting from the initialization of the simulation, $P_1$ is initialized to zero at the initial time. Rewriting Equation B1 for tracer $P_1$ in the fully coupled CO$_2$ quadrupling simulation, for example, we have

$$\frac{\partial P_1}{\partial t} = Q' - v \cdot \nabla P_1$$  \hspace{1cm} (B2)

The evolution of $P_1$ is thus equivalent to that of surface-driven temperature components in the FULL and PARTatm simulations (i.e., $P_1 = T'_{\text{OQ}}$ or $T'_{\text{OQ2}}$; compare Equation 5 or 8).
A second tracer denoted as $P_2$, tracks the temperature anomaly driven by circulation changes. $P_2$ emulates the evolution of a baseline temperature field $T_0$ (derived from an unperturbed simulation) driven by a perturbed ocean circulation ($v = v + v'$). The baseline temperature field in an unperturbed circulation or time-invariant circulation evolves according to the Equation A2:

$$\frac{\partial P_2}{\partial t} = Q - v \cdot \nabla P_2$$  \hspace{1cm} (B3)$$

Since $P_2$ evolves in a perturbed or time varying circulation, $P_2$ deviates from $T_0$ with time only because of the changes or variation in the circulation. Defining $P_2 = T_0 + T_{O_2}$, ($T_{O_2}$ is the deviation of $P_2$ from $T_0$) and rewriting Equation B3, we have

$$\frac{\partial (T_0 + T_{O_2})}{\partial t} = Q - (v + v') \cdot \nabla (T_0 + T_{O_2})$$  \hspace{1cm} (B4)$$

Taking the difference of Equations B4 and A2 produces

$$\frac{\partial T_{O_2}}{\partial t} = -(v' \cdot \nabla T - v \cdot \nabla T_{O_2})$$  \hspace{1cm} (B5)$$

Figure B1. Surface- and circulation-driven SST anomalies estimated using tracers $P_1$ and $P_2$ in the partially coupled simulation: (a) surface-driven component ($SST'_{OQO} = P_1$), (b) circulation-driven component ($SST'_{OOP} = (P_2 - T_0)$), and their sum ($P_1 + P_2 - T_0$), and the total surface temperature anomaly ($SST_p = T_0$).
Equation B5 is equivalent to the circulation-driven temperature anomaly evolution as defined in Equation 6 or (9), so that the difference between tracer $P_2$ and $T_O$ isolates the circulation-driven temperature anomaly in each simulation (i.e., $P_2 = T_O + T'_{OO}$ or $T'_{OO}$).

The total temperature in each simulation is thus equivalent to the sum of the two tracers (i.e., $T_O = T_O + T'_{OO} + T'_{OO}$). Although the tracers are not used in coupling in the FULL simulation, the ocean temperature passed to the coupler $T_{cpkl}$ is equivalent to $P_1 + P_2$, while in the PARTatm simulation, $T_{cpk}$ is $T_O + P_1$ (Table 1).

Lastly, we explain the tracer formulation the drives the PARTocn simulation (i.e., a variant of tracer $P_2$). In this case, tracer $P_2$ is initialized as the baseline temperature $T_O$ and also used for computing the surface flux exchange ($T_{cpk} = P_2$); $P_2$ is also forced with surface flux exchange (i.e., $Q_2 = Q_{cpk}$). Over time, tracer $P_2$ drives anomalous surface heat fluxes so that $Q_2 = Q_{cpk} = Q + Q'_{2}$. $P_2$ thus differs from the ones in the other experiment because $Q_2 = Q + Q'_{2}$ rather than just $Q$. $P_2$ therefore evolves according to

$$\frac{\partial P_2}{\partial t} = Q + Q'_{2} - v \cdot \nabla P_2 \tag{B6}$$

Over time, the deviation of this $P_2$ variant from $T_O$ also includes anomalies caused by $Q_2^e$ in addition to anomalies caused by circulation changes. $P_2$ after the initial time can be denoted $P_2 = T_O + T'_{O_2}$, where $T'_{O_2}$ includes two components $T'_{O_{21}} + T'_{O_{22}}$, driven by $Q_2$ and $v'$, respectively, so that Equation B6 becomes

$$\frac{\partial (T_O + T'_{O_{21}} + T'_{O_{22}})}{\partial t} = \nabla Q + Q'_{2} - (v + v') \cdot \nabla (T_O + T'_{O_{21}} + T'_{O_{22}}) \tag{B7}$$

Differencing Equations B7 and A2, we have

$$\frac{\partial (T'_{O_{21}} + T'_{O_{22}})}{\partial t} = Q_2^e - v' \cdot \nabla T_O - v \cdot \nabla T'_{O_{21}} - v \cdot \nabla T'_{O_{22}} \tag{B8}$$

Equation B8 can further be split into two parts:

$$\frac{\partial T'_{O_{21}}}{\partial t} = Q_2^e - v \cdot \nabla T'_{O_{21}} \tag{B9}$$

$$\frac{\partial T'_{O_{22}}}{\partial t} = -v' \cdot \nabla T_O - v \cdot \nabla T'_{22} \tag{B10}$$

Equation B9 is equivalent to the surface-driven temperature response to $Q_2^e$ by definition and (B10) is equivalent to the circulation-driven component. By definition $Q_2$ is equivalent to the ocean-driven anomalous surface fluxes $Q_O$ according to Equation 2.1, so that $P_2 = T_O + T'_{OO} + T'_{OO}$.

Note that the two tracers added to each perturbed simulation are independent of each other and the ocean temperature. A comparison of the surface temperature anomaly components isolated using the tracers and their sum with the total temperature anomaly (Figure B1) confirms that the decomposition is accurate.

**Data Availability Statement**

The model output for the control simulation data is available on the HPSS at NCAR (/CCSM/csm/b.e11.B1850C5CN.f09_g16.005). Perturbation simulation data can be download online (https://doi.org/10.5281/zenodo.3593507).

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