Dependence of global radiative feedbacks on evolving patterns of surface heat fluxes

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Abstract

In most climate models, after an abrupt increase in radiative forcing the climate feedback parameter magnitude decreases with time. We demonstrate how the evolution of the pattern of ocean heat uptake—moving from a more homogeneous toward a heterogeneous and high-latitude-enhanced pattern—influences not only regional but also global climate feedbacks. We force a slab ocean model with scaled patterns of ocean heat uptake derived from a coupled ocean-atmosphere general circulation model. Steady state results from the slab ocean approximate transient results from the dynamic ocean configuration. Our results indicate that cloud radiative effects play an important role in decreasing the magnitude of the climate feedback parameter. The ocean strongly affects atmospheric temperatures through both heat uptake and through influencing atmospheric feedbacks. This highlights the challenges associated with reliably predicting transient or equilibrated climate system states from shorter-term climate simulations and observed climate variability.

1. Motivation

The linear forcing-feedback framework assumes that the warming contribution of globally averaged climate feedbacks depend linearly on the global average near-surface temperature response following a radiative forcing, i.e., that the feedbacks are constant. The net global feedbacks are negative, counteracting the radiative forcing and stabilizing the global mean temperature. Many studies have assumed a constant global climate feedback parameter [e.g., Andrews et al., 2012; Forster et al., 2013; Otto et al., 2013], although some studies show that its magnitude decreases with time following an abrupt CO2 forcing [e.g., Senior and Mitchell, 2000; Gregory et al., 2004; Meraner et al., 2013; Andrews et al., 2015; Knutti and Rugenstein, 2015]. Here we focus on an open question with strong implications for the predictability of global climate evolution on time scales of decades to millennia: To what degree is the global climate feedback parameter dependent on the spatial patterns of surface temperatures and heat fluxes?

Ocean heat uptake, defined as net surface heat flux into the ocean, has a direct cooling effect on the atmosphere but indirectly affects surface temperatures through changing the magnitude of local radiative feedbacks. Thus, ocean heat uptake patterns might explain part of the changing magnitude of the global feedback parameter within one model, the spread between models, and the difference between models and observations [Winton et al., 2010; Bitz et al., 2012; Paynter and Frölicher, 2015; Gregory and Andrews, 2016; Rose and Rayborn, 2016].

Two studies in particular show that atmospheric feedbacks in idealized aquaplanets are very sensitive to the spatial structure of the ocean heat uptake [Rose et al., 2014] or ocean heat release [Kang and Xie, 2014]. Not primarily concerned with ocean heat uptake patterns, Andrews et al. [2015] show that 85% of models taking part in the Coupled Model Intercomparison Project Phase 5 (CMIP5) show a significant (22–62%) reduction in magnitude of the global feedback parameter in years 21–150 compared to years 1–20 following an abrupt4xCO2 forcing. Armour et al. [2013] argues that for one of these models, the reduction of the global feedback parameter comes about through the local structure of warming and ocean heat uptake enhancing constant local feedback strengths: High latitudes warming—where the magnitude of feedbacks are less negative, or positive, thus, destabilizing—leads to a substantially stronger global temperature response than low latitudes warming—where stabilizing, negative feedbacks are stronger.

Our aim now is threefold: First, we introduce a new experimental design to quantify transient feedback strengths through forcing a slab ocean model with a series of mixed layer depth heat fluxes derived from a
Figure 1. Slab ocean simulations forced with heat flux anomalies derived from a coupled simulation reproduce the temporal evolution of the feedback parameter. (a) Global top-of-the-atmosphere imbalance through a 2000 year long equilibration after an abrupt 4xCO₂ forcing for a large initial condition ensemble [Rugenstein et al., 2016]. Gray dots are individual ensembles' annual averages, black dots are ensemble mean annual averages (20 year running mean from year 150 onward). (b) Slope of the regression to the data in Figure 1a). Colored arrows indicate the different magnitudes of feedback parameters we reconstruct with slab ocean simulations, except for the “150 year regression” which is a linear regression over the first 150 years of data in Figure 1a). Arrows with black borders indicate simulations with sea ice; in all other simulations sea ice growth is inhibited. (c) Zonal averaged ocean surface heat flux anomaly through time in the abrupt 4xCO₂ coupled simulation. (d) Annual average heat flux average heat flux from the mixed layer to the interior ocean in the control simulation. Fluxes in Figures 1c and 1d are positive downward.

2. Models and Method: Generation of Q-Flux Forcing

We use the fully coupled (ocean-sea-ice-atmosphere-land) and the slab ocean (slab-ocean-sea-ice-atmosphere-land) configuration of the Community Earth System Model 1.0.4 with a finite volume horizontal resolution of 1.9° × 2.5° for the atmosphere and roughly 1° × 1° for the sea ice and ocean components [Bitz et al., 2012; Hurrell et al., 2013]. Figure 1a shows a large initial condition ensemble of the coupled model equilibrating the top-of-the-atmosphere (TOA) and surface temperature imbalances following an abrupt 4xCO₂ forcing. The slope of the data points is the global feedback parameter, and Figure 1b shows its 60% decrease over 200 years obtained through linearly regressing all annual means in a 1.3 K wide window which is moved in 0.1 K steps through the whole temperature range in Figure 1a (method explained by Rugenstein et al. [2016]). The TOA imbalance caused by the abrupt 4xCO₂ forcing is mostly mitigated by ocean heat uptake (Figure 1c), which moves within 10 years from a zonally near-homogeneous to a heterogeneous pattern. The mixed layer equilibrates in 10–20 years, after which the heat fluxes at the surface and the bottom of the mixed layer are roughly the same. In the slab ocean configuration, the heat flux at the bottom of the mixed layer (Q-flux) is prescribed and on average the same as the equilibrated surface heat flux, since there is no lateral heat
Figure 2. Heat flux imbalances of the same global magnitude but different spatial patterns result in different surface warming and feedback magnitudes. (top row) Q-flux patterns, all scaled to globally average to 2 W m\(^{-2}\) into the ocean. Idealized (a) high and (b) low latitude heat uptake, realistic decadal averages of the coupled run’s (c) first, (d) fifth, and (e) 20th decade. (middle row) Resulting equilibrium warming pattern, after applying both 4xCO\(_2\) (positive forcing, warming) and the Q-flux pattern (negative forcing, cooling, global value indicated above each panel), normalized by the global average equilibrium temperature change. (bottom row) Geographical distribution of the shortwave cloud radiative effect (SW CRE). Other feedback components are shown in Figure S6.

We apply six time invariant ocean heat uptake patterns (Figures 2a–2e): sine-shaped bands “Idealized high latitudes” and “Idealized low latitudes” to mimic Rose et al. (2014) and Kang and Xie (2014), “Idealized homogeneous” (not shown, uniform W m\(^{-2}\) anomaly), and three spatially varying “Realistic” patterns reconstructed from the coupled model (see section 3.2). “Realistic” here means imitating the coupled model behavior, not real world observations. The rationale is to analyze the coupled model’s behavior with a series of slab ocean simulations, each imitating one period in time of the coupled simulation (colored arrows in Figure 1b and discussed in section 4). To quantify the feedbacks associated with each ocean heat uptake pattern, we introduce a new approach: We scale the pattern up and down with a constant factor, to produce a global mean ocean heat flux of 1, 2, 3, and 4 W m\(^{-2}\). Each of these cases is a combination of positive 4xCO\(_2\) and negative ocean heat uptake forcing of varying strength and pattern. All changes in the feedback parameter here are solely due to different ocean heat uptake patterns as opposed to temperature or time dependencies or Earth system transport within the slab. The partitioning between the shortwave (SW) and longwave (LW), sensible and latent heat fluxes, and the sea surface temperatures (SST) evolve freely within the prescribed Q-flux constraint.
feedbacks [e.g., Senior and Mitchell, 2000; Meraner et al., 2013; Knutti and Rugenstein, 2015; Gregory et al., 2015]. Throughout the paper we show the final decade of an equilibrated 40 year long simulation. The feedbacks are quantified through linearly regressing the top-of-atmosphere imbalance (which in equilibrium is equal to the prescribed net ocean heat uptake) on the near-surface change in air temperature. Note that there is virtually no internal variability, and the uncertainty on the regression is thus very small (supporting information Table 1). We also circumvent the problem of regressing points which are not equally spread in the temperature space. The approach is similar to the SST pattern scaling by Andrews et al. [2015], but we—in addition to using a different model—constrain only the net ocean heat uptake, while the SST and heat flux components can evolve freely. Thus, we link the changing feedback parameter to heat uptake and not to changing SST patterns only. Our approach differs from the 2 W m$^{-2}$ imbalance of Rose et al. [2014] and the 3.3 W m$^{-2}$ imbalance of Kang and Xie [2014] in that we use realistic topography and patterns of ocean heat uptake, spatially varying mixed layer depth, sea ice, seasonal varying solar insolation, a large range of flux imbalances, a different reference state, and a new method to quantify feedbacks. Our aim here is to explain the typical coupled climate model behavior and potentially real world climate evolution. We analyze to what extent idealized setups mimic realistic behavior and thus close an important gap in the model hierarchy. To compare to the idealized studies done without sea ice and due to slab ocean configuration technicalities, we simulate all cases without and some cases with sea ice (see details on allowing sea water to supercool in the supporting information). Limitations of our setup are the use of a single model (though according to Rose et al. [2014], it is representative of at least three other models) and the strong sensitivity of sea ice growth to the local Q-fluxes [Rose, 2015]. Note that the Q-fluxes are not symmetric about the equator in most cases, shifting the intertropical convergence zone, which is not our focus here, but itself an area of research [e.g., Kang et al., 2008, 2014; Zhang et al., 2010].
3. Results

We first discuss the influence of the idealized ocean heat uptake forcing on global and regional temperatures, the global net feedback, its different components, and their geographical distribution. We then show the same metrics for the realistic cases and discuss how they differ from the idealized ones.

3.1. Temperature and Feedback Responses to Idealized Pattern

Figures 2f and 2g show the warming pattern for two of the idealized cases normalized with their global equilibrium temperature and confirm that the overall magnitude and the spatial structure of the warming sensitively depend on the ocean heat uptake pattern. Above each panel, the cooling induced by the ocean heat uptake forcing relative to the warm reference state is indicated. High-latitude ocean heat uptake is thus 3.5 times (3.3 without sea ice) more effective in cooling the atmosphere than low-latitude ocean heat uptake of the same amount (here, globally, 2 W m$^{-2}$). In other words, the ocean heat uptake efficacy—introduced by Winton et al. [2010] to compare the radiative effect of CO$_2$ alone (in our case 1.1 W m$^{-2}$ K$^{-1}$ for the transient response of abrupt4xCO$_2$ to the control Q-flux pattern) to the radiative effect of ocean heat uptake—is small for the low-latitude case ($\epsilon_{\text{low lat}} = 0.48$) and large for the high-latitude case ($\epsilon_{\text{high lat}} = 1.57$). This compares well with Rose et al. [2014], but is substantially smaller than Kang and Xie [2014]—who find 13 times stronger effects of high- versus low-latitude heat release, prescribed in a narrower region.

Figure 3a shows the global feedback parameter for the three idealized cases. Every dot is a decadal average of an equilibrated slab ocean simulation with different global imbalances for each ocean heat uptake pattern. Solid lines and filled dots indicate simulations that do include sea ice, dashed lines and open circles, those that do not. The regression is a linear least squares fit through all available points (four to six for each ocean heat uptake pattern, including the reference state at (0,0)). The horizontal axis shows the cooling relative to the reference state. Overall, for each heat flux pattern, the feedbacks are linear for the temperature and heat flux range tested here. The sea ice may or may not, depending on the ocean heat uptake pattern, amplify the response: for the high-latitude case, the sea ice response makes up 33% of the total response (similar to Caldeira and Cvijanovic [2014]), while for the homogeneous and low-latitude case it is only 11 and 3%.

Figure 3c shows the different feedback components for each case. The LW clear sky (red) feedback is more negative in the low-latitude case but otherwise does not change much, since we do not cover a large range of temperatures [Meraner et al., 2013]. The SW clear sky (blue), reflecting the surface albedo, differs most strongly between the cases with and without sea ice. The LW cloud feedback (white) is generally small but changes sign between the low- and high-latitude cases. The largest contribution to the smaller magnitude of the global feedback parameter in the high- versus low-latitude case arises from the SW cloud radiative effect (SW CRE, black). Figure 2 (bottom row) shows its geographical distribution. In each grid box four equilibrated end points of the scaled pattern simulations are regressed against the global temperature anomaly. The feedback pattern is completely different in most regions and sensitively depends on the sea ice formulation for the low-latitude case. Compared to the Rose et al. [2014] aquaplanet study, the signs of the global feedbacks are the same, but their magnitudes are substantially smaller in our idealized setup. In the supporting information we discuss other feedback components (Figure S6), the simulations without sea ice (Figure S5), and cloud masking effects (Figures S3 and S4), which do not impact our conclusions.

3.2. Comparison to Realistic Patterns and Ocean Heat Release

In the coupled simulations, the heat uptake patterns differ from the idealized in many aspects. The three representative realistic Q-flux patterns (Figures 2c–2e) feature the “Realistic 1st decade” with a relatively homogeneous ocean heat uptake, the average response of the “Realistic 5th decade”, in which the reduction of the Atlantic Meridional Overturning Circulation (AMOC) is strong (red patch in the North Atlantic) and the Southern ocean heat uptake becomes important, and finally, the “Realistic 20th decade”, in which the AMOC has restrengthened and the Southern Ocean becomes the dominant heat sink [e.g., Frolicher et al., 2014; Li et al., 2013]. These latter two patterns are representative of CMIP5 models surface heat flux after 100 years [Marshall et al., 2014]. We cannot differentiate which part of the pattern is due to the atmospheric or oceanic influence only [e.g., Stouffer and Manabe, 2003; Xie et al., 2010; Exarchou et al., 2014], since we deduce the Q-flux from the coupled model. In parts of the tropics but also high latitudes, the heat uptake capacity of the local ocean saturates, and the circulation shifted enough to make some places strong heat sources for the mixed layer and the atmosphere. Since the heat fluxes are locally large enough to lead to excessive sea ice growth, we simulate the whole suite of realistic experiments without sea ice for cleaner comparison.
High-latitude surface heat flux imbalances result in a far-field effect and large meridional heat transport. Zonally averaged top-of-the-atmosphere imbalance, atmospheric heat transport divergence (convergence positive), and heat flux into the interior ocean — relative to the 4xCO₂ equilibrium for Idealized low latitude pattern (yellow shading and arrows) and high latitude pattern (red) each averaging globally to a 3 W m⁻² imbalance. Arrows indicate local and far-field balancing of the surface perturbation. Figure S8 shows realistic cases.

Figures 2h–2j show again the normalized temperature response to a 2 W m⁻² ocean heat uptake forcing in a 4xCO₂ climate. The first decade pattern induces a relatively homogeneous warming, while stronger local heat uptake leads to a small local temperature response (white patches in Figures 2i and 2j). The polar amplification is reduced (Figures 2h and 2i) compared to the idealized cases. However, a strong Arctic amplification occurs in the 20th decade, in which the realistic coupled models’ sea ice is almost completely melted, causing large upward heat fluxes into the mixed layer. Note that the ocean heat uptake can influence land surface temperatures on all continents even centuries after the application of the abrupt 4xCO₂ forcing and shift the ocean-land warming contrast. Absolute temperatures are not shown here, but land warming is stronger in India and South Africa and weaker in Europe in the 20th than in the 5th decade, even though the global temperature is smaller.

Figures 3b and 3c show again the global feedback parameter and its components for each case. In contrast to the idealized cases in the realistic cases the LW cloud feedback explains more of the difference between the cases (see Table S1) and is the only feedback changing sign. The spatial SW CRE distribution (Figures 2m–2o and supporting information Figure S7) does — in most regions — only change its magnitude. Exceptions are the eastern tropical Pacific, the North Atlantic, and the Southern Ocean. This comparably subtle change in local feedback (compares Figures 2k versus 2l with Figures 2m versus 2o) — which still aggregates to a strong global signal — might be the reason why Armour et al. (2013) found a description of locally constant feedbacks suitable. In agreement with the idealized setup that the SW CRE contributes strongest to the difference between the cases.

We reiterate the reason for the different temperature response from the heat flux perspective with the help of an illustrative figure. Figure 4 depicts the heat fluxes at the surface, the TOA, and their difference — the local divergence — at each latitudinal band for the idealized cases of low-latitude (yellow) and high-latitude (red) ocean heat uptake, relative to the warm reference state. The low-latitude ocean heat uptake is balanced mostly by the TOA fluxes within the same latitudinal bands (yellow range between 30°S and 30°N); thus, meridional heat transport is necessary to compensate the exact regional surface heat flux pattern, but occurs within the latitudes of the prescribed anomaly. Because the atmosphere is unstable, the surface anomalies readily reach tropopause height, and the outgoing LW radiation strongly depends on the temperature (see more details in Kang and Xie [2014]). The result is a weak global cooling (Figure 2g). For other shapes of surface heat flux forcings the meridional heat transport into the tropics can be larger and caused by the suppression of subtropical evaporation and poleward latent heat transport [Rose et al., 2014]. The high-latitude ocean heat uptake, however, is not balanced locally at the TOA due to the stably stratified atmosphere (red-shaded areas and arrows). Large heat transport into the regions of the ocean heat uptake are necessary and lead to a stronger polar amplified warming compared to the low-latitude case. Thus, high-latitude ocean heat uptake has stronger far-field effects and is more efficient in changing global temperatures than low-latitude ocean heat uptake. Figure S7 confirms this, showing the geographical distribution of each feedback component: The case with the stronger high-latitude heat flux (“Realistic 20th decade”) has a stronger influence on the low-latitude
cloud feedbacks than the case with a less amplified high-latitude pattern (“Realistic 5th decade”). However, with our setup we cannot tell whether the cloud response causes or reacts to the meridional heat transport. This highlights the necessity of studying the cause and effect of global ocean surface flux pattern, meridional heat fluxes, and local cloud responses [e.g., Rose and Rayborn, 2016; Trossman et al., 2016].

The heterogeneity of ocean heat uptake, even to the point of local heat release (blue in Figures 1c–1e) is little appreciated so far in the literature. The Q-flux patterns differ little in the regions of heat uptake between the 5th and 20th decade (the Nordic Seas being an exception), but the magnitude of the local surface fluxes differ at smaller spatial scales. For example, the Southern Ocean takes up a lot more heat at later stages during the equilibration, but it also releases large amounts a few degrees farther north and south. In reality and coupled models, the dominant mixing length scale of the surface ocean controls how strong the heterogeneity of (deep) ocean heat uptake imprints on the SSTs and surface heat flux patterns. Further studies on heat uptake and release and their relative, local, and far-field impact are needed to determine whether the effect found here in the slab ocean setting carries over to the real world. In the supporting information we discuss other spatial homogeneous and heterogeneous patterns as a first sensitivity test.

Overall, our results show for the first time that the findings from idealized cases do carry over to realistic cases, with three exceptions. First, strong low-latitude focused ocean heat uptake or release seems not to occur without changing pattern elsewhere in a decadal averaged coupled system. Consequently, the difference in feedback components between two idealized cases is more than twice as large as the differences between two realistic cases. Next to the SW CRE, the difference between two cases is set mostly by the SW clear sky feedback for the idealized cases and by the LW CRE for the realistic cases (Figure 3c and Table S1). The magnitude of feedbacks is generally smaller in the realistic cases. Second, sea ice may be—depending on the spatial Q-flux forcing pattern warming—an important part of the overall feedback response. Third, the ocean heat release and the exact pattern and distance of release and uptake might play an important role in the noisy real world.

4. Implications and Outlook

We have shown that different patterns of SST and surface heat fluxes—induced and constrained by heat fluxes from the mixed layer into the deep ocean—can lead to a continuous time evolution of the global climate feedback parameter in a coupled model, which is usually assumed to be constant. For each specified ocean heat flux (Q-flux) pattern, global climate feedbacks are linear in the temperature and heat flux range we test, but the magnitude of the feedback parameter and its components depend strongly on the Q-flux pattern.

Figure 1b compares the global feedback parameter of all Q-flux forcing cases (colored arrows) to the one fully coupled simulation. Because the slab ocean simulations include some mixed layer averaging, a fixed mixed layer depth, a different reference state, and in some cases no sea ice, the feedback parameters do not match the specific “realistic” decade they represent [Shell, 2013]. Overall, however, the cases mimic the evolution of homogeneous to high-latitude ocean heat uptake. Additionally, we show that the homogeneity of the ocean heat uptake pattern can explain variations in the global feedback parameter of the coupled system. The 150 year regression (blue arrow, calculated by regressing the coupled transient response of the first 150 years following the abrupt4xCO₂ forcing) results in an arbitrary feedback parameter [Andrews et al., 2012, 2015] and the low-latitude ocean heat uptake cases result in an overly unsensitive state never reached in the coupled model’s reality. As a sideline, our analysis confirms that slab ocean simulations with a prescribed constant climatological Q-flux might not be the preferred tool to study transient behavior of atmosphere and land [e.g., Jonko et al., 2013; Bitz et al., 2012; Shell, 2013; Donohoe et al., 2014; Deser et al., 2015], even if global mean ocean heat uptake is negligible.

Resulting follow-up questions include the following: Which local features and physical mechanisms of the heat uptake in which geographical combination are most efficient in changing the magnitude of the global feedback parameter and are we sure they will occur in reality and are not modeling artifacts [e.g., Zhang et al., 2010; Exarchou et al., 2014; Liang et al., 2015]? What is the critical size of a region with a certain ocean heat uptake to influence the magnitude of the global feedbacks [L’Hévéder et al., 2015; Kang and Xie, 2014, and Figures S9 and S10]? How do decadal variability, the time scales of changing patterns, and the superposition of different forcings (aerosols versus ocean heat uptake) modulate our findings and the transferability to the real world [Hansen et al., 1997; Ban-Weiss and Caldeira, 2010; Hsieh et al., 2013; Long et al., 2013; Gregory and Andrews, 2016; Dallafior et al., 2016]? When and where does local ocean heat uptake saturate and anomalous heat release become important relative to the climatological heat fluxes in realistic scenarios.
Subtle changes in wind stress or ocean circulation might have a small direct influence in terms of net heat uptake but a large indirect impact on surface temperature, through modifying feedback magnitudes. How much of polar amplification and its evolution is solely due to ocean heat-flux-pattern-induced atmospheric heat flux convergence? What is the physical mechanism of ocean heat uptake and cloud response and how does it evolve in the coupled system [Rose and Rayborn, 2016; Trossman et al., 2016]? How much of the disagreement among models in simulated feedback strengths is due to the difference among the models in ocean heat uptake patterns [Winton et al., 2010] or ocean heat-uptake-induced SST changes [Long et al., 2014]? The use of a constant global feedback parameter in intermediate complexity models, prediction, and impact studies should be—depending on the purpose of use—carefully considered and may turn out to be problematic. The ability to predict transient and equilibrium behavior on any time scale beyond a decade critically depends on our understanding, regional observations, and ability to correctly simulate SST pattern formation, ocean heat uptake pattern, and the atmospheric response to those changing fluxes—as they evolve through time.

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