

# Feedbacks and Eddy Diffusivity in an Energy Balance Model of Tropical Rainfall Shifts

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## Abstract

Radiative feedbacks are known to strongly modify horizontal shifts of the intertropical convergence zone (ITCZ) produced by remote atmospheric energy sources. This study uses a one-dimensional moist energy balance model to understand how radiative feedbacks and the structure of an imposed eddy diffusivity can influence such ITCZ shifts. The Planck feedback is shown to damp ITCZ shifts about twice as strongly for extratropical forcings as for tropical ones, because lower moisture content in cold regions makes the temperature response larger there. The water vapor feedback on ITCZ shifts is shown to be dominated by changes in the cross-equatorial asymmetry of the relative humidity of subtropical dry zones, with additional contributions by the changes in mixing ratio that occur at fixed relative humidity and by the meridional shift of the humid ITCZ. Finally, the ITCZ response is found to be highly sensitive to the meridional structure of the diffusivity; the ITCZ shift increases with the tropical diffusivity, even when the global mean diffusivity is fixed.

Atmospheric circulations transport both water and energy, coupling the spatial structure of precipitation to the spatial structure of atmospheric energy transport. When the measure of energy is chosen to include the latent heat stored in water vapor, the net input of this moist energy through the top and bottom of the atmosphere can then be related to the position and intensity of time-mean tropical precipitation maxima, without requiring specification of the latent heat release that accompanies precipitation [1]. Such moist energy frameworks have improved understanding of a variety of tropical climate phenomena, including tropical cyclones, monsoons, and the Madden-Julian Oscillation [see reviews by 2, 3, 4].

A salient example of the improved understanding resulting from moist energy frameworks is the recognition that the latitude of the intertropical convergence zone (ITCZ) can be influenced by high-latitude anomalies in the atmospheric moist energy source [e.g. 5, 6]. Specifically, the zonal mean ITCZ shifts toward an anomalous energy source, in the opposite direction of the cross-equatorial, vertically integrated atmospheric energy flux needed to maintain a steady-state energy balance in the presence of such a forcing [7, 8]. This ITCZ shift can be quantitatively diagnosed from the anomalous energy sources and fluxes, through either empirical fits [9] or Taylor expansions for the latitude of the energy flux equator (EFE), defined as the zero line of the zonal mean atmospheric meridional energy transport [10, 11]. Related quantitative frameworks have also been used to go beyond the zonal mean and provide information on regional precipitation shifts [12, 13, 14]. These studies that relate horizontal shifts in tropical precipitation to anomalous energy sources do not require knowledge of the vertical structures of winds and moist energy content, unlike some of the original moist energy-based frameworks [1, 15]; they correspondingly provide no information on changes in the amplitude of the precipitation field.

Most frameworks that relate horizontal shifts in precipitation to anomalous sources of atmospheric moist energy are furthermore diagnostic rather than prognostic: they require knowledge of the net energy source anomaly in order to provide information on the precipitation shift. This is because feedbacks act on any imposed energy source, producing changes in radiative fluxes and in surface turbulent fluxes of latent and sensible heat, altering the horizontal at-

mospheric energy transport needed to balance the imposed forcing [e.g. 16, 8, 17]. These feedbacks can be large. For example, dynamical ocean heat transports are estimated to damp the zonal mean ITCZ shift by a factor of three compared to what would be achieved in the absence of a dynamical ocean [18, 19]. Water vapor provides a positive feedback on the ITCZ shift, consistent with the fact that it provides a positive feedback on the global mean temperature response to a global mean radiative forcing [20, 21, 22]. Clouds can provide a positive or negative feedback on ITCZ shifts, at least in numerical models with varying representations of cloud properties [23, 24], with the sign of the feedback likely depending on whether the opposing longwave or shortwave cloud feedback dominates [see discussion by 20].

Here we use a diffusive energy balance model to improve understanding of the influence of radiative feedbacks on ITCZ shifts produced by localized radiative forcings. We are motivated by the fact that the influence of individual feedbacks on ITCZ shifts remains poorly understood, in part because climate feedbacks have been studied most extensively in the context of the global mean temperature response to global mean radiative forcings [e.g. 25, 26, 27]. This is not to say that radiative feedbacks on ITCZ shifts have been unexplored; global climate models (GCMs) with idealized and realistic boundary conditions have been used to study the influence of cloud, water vapor (WV), albedo (AL), and lapse rate (LR) feedbacks on the ITCZ response to imposed forcings [8, 17, 23, 24].

One systematic exploration of an individual feedback was performed by Clark et al. [22] (hereafter C18), who used an idealized GCM to assess the influence of the WV feedback on ITCZ shifts produced by radiative forcings of different magnitudes located in either the tropics or extratropics. Unlike prior studies that diagnostically estimated the strength of individual feedbacks using a classic climate sensitivity framework, yielding coefficients for feedback strengths [e.g. 28], C18 used prognostic models (GCMs) with and without the WV feedback to show that this feedback roughly doubled the ITCZ shift produced by a given forcing.

Here we follow a similar methodology, but we use a one-dimensional (1D) moist energy balance model (MEBM) like that employed in studies of the influence of anthropogenic greenhouse gas forcing on poleward energy transports [29], polar amplification of warming [30], and regional climate pre-

dictability [31]. Because of the computational efficiency of the MEBM, this allows us to prognostically quantify the influence of multiple feedbacks on ITCZ shifts; the diffusive nature of the model also allows us to assess the comparative importance of diffusive-like transports in different regions. We view this as a small first step toward a theoretical framework that may eventually be used to quantitatively predict feedbacks on ITCZ shifts.

## RESULTS

Here we present results from both an MEBM and a GCM. The MEBM represents the steady-state, vertically integrated moist energy budget in sine-latitude coordinates ( $x = \sin \phi$ ) as

$$-\frac{\partial}{\partial x} \left[ D(x)(1-x^2) \frac{\partial h}{\partial x} \right] = NEI. \quad (1)$$

The left-hand side represents the vertically integrated divergence of moist energy by the circulation as meridional diffusion of the surface air moist static energy  $h$  with diffusivity  $D(x)$ . The net energy input to each atmospheric column is  $NEI = S(1 - \alpha) - L$ , where  $S$  is insolation,  $\alpha$  is albedo, and  $L$  is top-of-atmosphere (TOA) outgoing longwave radiation. The MEBM uses a sophisticated radiative transfer scheme to represent the effects of temperature and water vapor on long-wave radiation. We use this MEBM together with a GCM (the Community Earth System Model version 2 [CESM2]) to conduct a series of simulations of the ITCZ shifts produced by remote forcings (see Methods section for more details). We run CESM2 as an entirely oceanic “aquaplanet” with a 1 meter-deep ocean mixed layer. Neither model has a representation of cloud-radiative effects.

## Forcings and responses

We apply a series of insolation forcings ( $S'$ ) to both the MEBM and CESM2, taken directly from C18. These are Gaussian energy sinks centered in the northern hemisphere (Fig. 1a). We conduct one set of simulations with perturbations centered at  $15^\circ\text{N}$  (labeled “tropical”) and another with perturbations centered at  $60^\circ\text{N}$  (“extratropical”). Each of these sets contains four simulations with different forcing magnitudes, defined as  $-M = \overline{S'}$ , with an overbar representing a global mean so that  $-M$  represents the prescribed global-mean change in insolation. For a given value of  $M$ , the tropical and extratropical forcings thus produce the same global-mean change in insolation. Figs. 1b, c show the MEBM’s temperature and MSE solutions for each forcing.

To assess the response to these forcings, we find the latitude of the EFE in each simulation. There is no explicit precipitation in the MEBM, and thus no explicit ITCZ, but the latitude of the EFE,  $\phi_E$ , has been shown to be proportional to the latitude of the ITCZ,  $\phi_I$  [e.g. 8]. For the simulations in C18, it was found that  $\phi_I = 0.64\phi_E$ . For our CESM2 runs, we found that  $\phi_I = 0.78\phi_E$  with  $r^2 = 0.996$ . Because the relation between EFE and ITCZ latitudes seems model-dependent, we analyze EFE shifts and do not explicitly relate these to ITCZ shifts.

## Overview of energy flux equator sensitivities

For both tropical and extratropical forcings, the control MEBM has an EFE latitude that exhibits a roughly similar sensitivity to the magnitude of the forcing as both CESM2 and the GCM run by C18 (Fig. 2, compare closed black circles to open gray circles connected by lines). The match between the MEBM and the GCMs is especially close for extratropical forcings, though the sensitivity to those forcings is only about one-third as large as the sensitivity to tropical forcings, consistent with previous model simulations performed without cloud

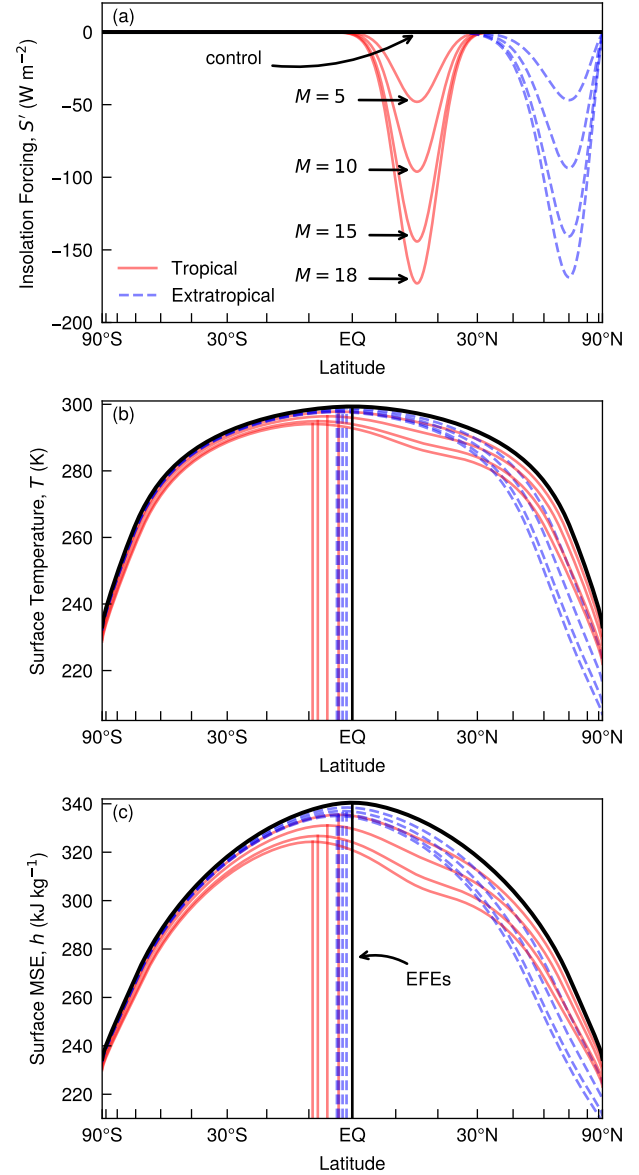


Figure 1: The insolation forcings used in sensitivity experiments and examples of the temperature and MSE solutions obtained by the MEBM. Tropical forcings and responses are in solid red while extratropical are in dashed blue. **(a)** Insolation forcings,  $S'$ , for strengths  $M = 5, 10, 15$ , and  $18$ , including control of  $M = S' = 0$ . **(b)** Surface air temperature,  $T$ , and **(c)** surface air MSE,  $h$ , distributions resulting from MEBM simulations with constant diffusivity,  $D$ , and AL, LR, WV, and PL feedbacks. Vertical lines mark the position of EFEs.

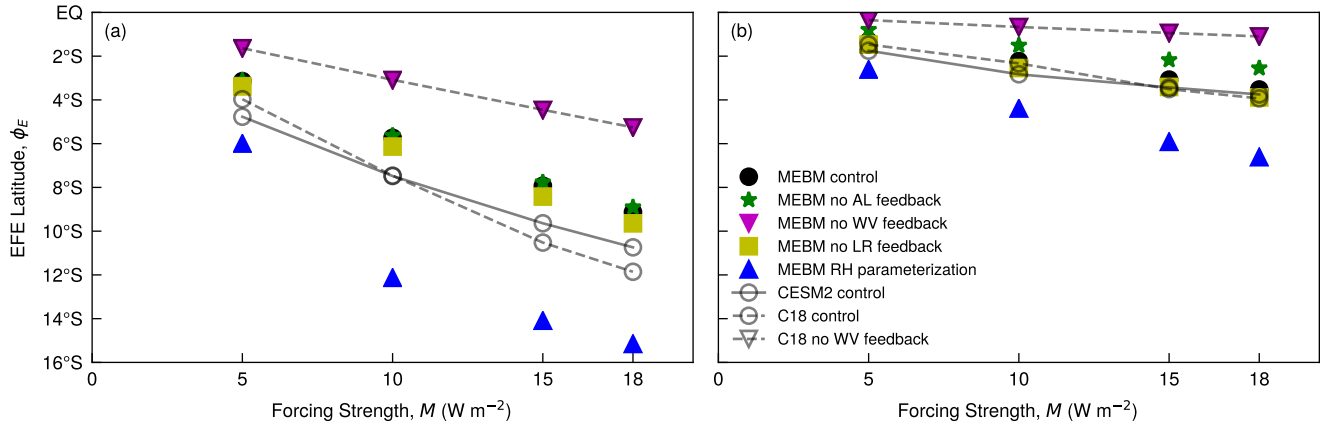


Figure 2: EFE shift,  $\phi_E$ , versus forcing strength,  $M$ , for GCM and MEBM simulations. **(a)** Tropical forcings. **(b)** Extratropical forcings. Black circles show sensitivity of MEBM with AL, WV, and LR feedbacks active. Other symbols refer to suppressed feedback simulations.

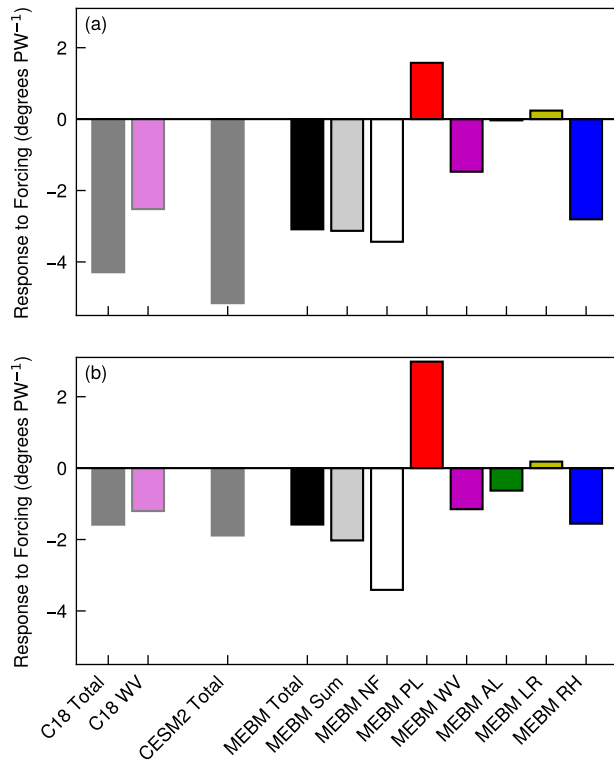


Figure 3: Linear responses of models to forcings. Calculated using data from **(a)** tropical and **(b)** extratropical forcing simulations. The linear response is defined in equation (2). Here we use only  $M = 5$  simulations to reduce nonlinearities. “Total” refers to simulations with all feedbacks included; bars marked with a particular feedback (e.g. WV) represent the strength of that individual feedback calculated using the difference between the all-feedback simulation EFE and suppressed feedback EFE. “MEBM Sum” represents the sum of PL, WV, AL, and LR bars and should match the total in a linear system. “MEBM NF” is the MEBM with all feedbacks turned off (no AL feedback and constant OLR at all grid points). The PL feedback response is calculated as a difference between simulations with no WV, LR, or AL feedbacks and the NF simulations.

feedbacks [23, C18]. The EFE latitude exhibits some nonlinearity in its relation to forcing amplitude even in the MEBM, but this nonlinearity is greater in the GCMs. For example, CESM2 exhibits a 4.8° EFE shift for a tropical forcing with  $M = 5 \text{ W m}^{-2}$ , while tripling the magnitude of that forcing only doubles the EFE shift. We discuss reasons for this nonlinearity below, but first we assess linear estimates of these sensitivities and their dependence on feedbacks.

We define the linear response of the EFE to a forcing in terms of the coefficient,  $\lambda$ , that relates the EFE latitude to the cross-equatorial atmospheric energy transport needed to balance the imposed forcing in the absence of all feedbacks. To define that cross-equatorial transport, we follow Kang et al. [8] and meridionally integrate the forcing after subtracting its global mean value, with the forcing defined as the net absorbed shortwave in the absence of any AL feedback,

$$\mathcal{T}_{S'(1-\alpha)}(x_0) = 2\pi a^2 \int_{-1}^{x_0} [S'(1-\alpha) - \overline{S'(1-\alpha)}] dx$$

where  $a$  is Earth’s radius and  $x_0$  is the sine of latitude at the basic state EFE (here equal to zero). The meridional atmospheric energy transport,  $\mathcal{T}$  (in PW), is the zonal integral of the meridional vertically integrated atmospheric energy flux,  $F$  (in  $\text{W m}^{-1}$ ). The forcing, expressed in terms of its cross-equatorial energy transport, is then  $\mathcal{T}_{S'(1-\alpha)}$  and the linear sensitivity to that forcing is

$$\lambda = \frac{\phi_E}{\mathcal{T}_{S'(1-\alpha)}}. \quad (2)$$

We calculate  $\lambda$  using the smallest forcing ( $M = 5 \text{ W m}^{-2}$ ) to better approximate the linear response. This provides sensitivities between about  $-1.5$  and  $-5^\circ \text{PW}^{-1}$  for the control MEBM and GCMs (see bars marked “Total” in Fig. 3). Although  $\lambda$  has units of  $^\circ \text{PW}^{-1}$ , it is distinct from the sensitivity estimate of  $-3^\circ \text{PW}^{-1}$  obtained by Donohoe et al. [9] based on climate models and observations, because Donohoe et al. [9] used the net cross-equatorial energy transport produced by the forcing and all feedbacks, while  $\lambda$  references the EFE shift to the forcing alone. This choice of  $\lambda$  also accounts for the fact that the albedo is higher in the polar regions of the basic state, giving the effective extratropical forcing a smaller global mean value than the tropical forcing (long-dashed lines in Fig. 4a, c); even after accounting for this effect, the sensitivity of the EFE to tropical forcings is still about twice the sensitivity to extratropical forcings.

We now estimate the influence of each feedback on the sensitivity of the EFE to the forcings, comparing the response

of the EFE in the control MEBM with the response in MEBM simulations in which individual feedbacks were suppressed. Removing the WV feedback strongly decreases the sensitivity of the EFE to both tropical and extratropical forcings, by amounts similar to those seen in the GCM simulations of C18 (Fig. 2, closed purple and open gray triangles). The LR feedback is comparatively weak and, in contrast to the WV feedback, is negative (i.e. increased EFE sensitivity is seen when the LR feedback is suppressed; yellow squares in Fig. 2). The AL feedback (green stars in Fig. 2) has almost no effect on the response to tropical forcings, whereas it provides a positive feedback on extratropical forcings with about half the strength of the WV feedback. In the MEBM, feedbacks on EFE latitude thus have the same sign as they do when altering the response of global mean surface temperature to a global mean radiative forcing [e.g. 27]. The WV feedback has a roughly similar strength for tropical and extratropical forcings, as does the LR feedback, while the AL feedback is active primarily for extratropical forcings.

These results can be presented succinctly by calculating the contribution of each feedback to the net linear sensitivity parameter  $\lambda$ . That is,  $\lambda = \lambda_{NF} + \sum_i \lambda_i$ , where  $\lambda_i$  is estimated for each feedback (AL, WV, LR, and PL), and  $\lambda_{NF}$  is the “no feedback” sensitivity obtained by running the MEBM with all feedbacks turned off (even PL) and with the forcing rendered energetically neutral in the global mean by subtracting its global mean from its value at each latitude. For example,  $\lambda_{AL}$  is estimated by subtracting the EFE shift in the run with the AL feedback suppressed from the EFE shift in the control run, and normalizing this difference by  $T_{S'(1-\alpha)}$ . The estimate for each  $\lambda_i$  is obtained from the corresponding run with the  $M = 5 \text{ W m}^{-2}$  forcing, as was done when estimating the net  $\lambda$ .

For tropical forcings,  $\lambda \approx \lambda_{NF}$ , but this seems to be a coincidence due to the approximate cancellation of the PL and WV feedbacks (Fig. 3a). The sum of  $\lambda_{NF}$  and all of the  $\lambda_i$ 's is very close to the net  $\lambda$  estimated from the control run, indicating that the feedbacks behave near linearly for tropical forcings. For extratropical forcings, the PL feedback is much larger and the net  $\lambda$  is only about half as large as  $\lambda_{NF}$  (Fig. 3b). Furthermore, the sum of the forcing and each individual feedback is about 20% larger than the estimate of  $\lambda$  obtained directly from the control run, suggesting the existence of nonlinearity in the feedbacks or their interaction for extratropical forcings (even the  $M = 5 \text{ W m}^{-2}$  forcing has a local insolation anomaly that peaks near  $50 \text{ W m}^{-2}$ , so the presence of nonlinearities here is not surprising).

## Spatial structure of feedbacks

We now examine the spatial structure of each feedback, decomposing the change in net energy input at each latitude into components due to the forcing and each individual feedback. We then relate each feedback to a component of the anomalous energy transport using the steady-state energy balance,

$$NEI' = \nabla \cdot F' = \nabla \cdot (F_{S'(1-\alpha)} - F_{(S+S')\alpha'} - F_{L'})$$

where we expressed the anomalous flux  $F'$  in terms of components whose divergence balances the sum of the forcing, the AL feedback, and all longwave feedbacks (as above, components of the anomalous transport  $T'$  are obtained by zonally integrating the corresponding component of the anomalous flux  $F'$ ).

We furthermore decompose the net change in OLR,  $L'$ , into components associated with each feedback, which we obtain from offline radiative transfer calculations. These offline calculations provide another assessment of the feedbacks that supplements the prognostic, suppressed-feedback simulations described above. To calculate  $L'_{WV}$ , for instance, we compute OLR with the perturbed simulation's temperatures but the control simulation's specific humidities, and subtract

this from the OLR in the perturbed simulation. The Planck (PL) feedback is diagnosed by computing OLR for a model state consisting of the forced simulation's water vapor profile and lapse rates, but with surface temperatures matching those of the control simulation. Summing the  $L'_i$  distributions produced by these offline radiative transfer calculations produces a good approximation to the  $NEI'$  diagnosed by differencing the perturbed and control models, confirming that the feedbacks can be diagnosed in this way and linearly superimposed, even for the example of the strong  $M = 15 \text{ W m}^{-2}$  forcing (Fig. 4a, c). That said, this analysis does provide evidence for some nonlinearity in the response to the extratropical forcing arising from feedbacks in the northern high-latitudes (cf. solid and dashed black lines in Fig. 4c).

The flux form of the feedbacks improves understanding when one recalls that a northward anomalous flux at the location of the basic-state EFE corresponds to an anomalous southward shift of the EFE [e.g. 9, 10]. For instance, the contribution of the AL feedback to  $NEI'$  is very localized compared to that of the WV feedback, but both have similar order-of-magnitude effects on the equatorial  $T'$  for extratropical forcings (Fig. 4d). For tropical forcings, the AL feedback has a small effect on the equatorial  $T'$  and thus a small effect on EFE shifts because its effect on the  $NEI'$  has a substantial equatorially symmetric component and also because its northern hemisphere contribution to the  $NEI'$  is located further poleward than that for extratropical forcings. Perhaps the most remarkable feature of the  $T'$  distributions is that the PL feedback has the largest contribution to the equatorial  $T'$ , and that contribution is furthermore twice as large for the extratropical forcing as for the tropical, consistent with our earlier estimates of  $\lambda_{PL}$ . We discuss the implications of these features of the PL feedback in the next subsection, and the nature of the WV feedback in the following subsection. As all of these estimates of  $T'$  assume diffusive energy transports, we examine the sensitivity to the assumed diffusivity in the penultimate section.

## Why extratropical forcings cause smaller EFE shifts

One of the most obvious questions one can ask about the results in Fig. 2 is why extratropical forcings produce such small EFE shifts compared to tropical forcings. Seo et al. [23] suggested that, without radiative feedbacks, the diffusive nature of energy transports causes the influence of forcings on EFE latitude to become smaller the further the forcings are from the equator (they also found that a strong cloud feedback can amplify the effects of extratropical forcings to become larger than those of tropical forcings, but neither our models nor that of C18 employ cloud feedbacks). We argue that the larger sensitivity of EFE latitude to tropical forcings is instead due to the fact that the PL feedback will be smaller for forcings in the warmer part of the planet. To see this, consider the steady, perturbation form of equation (1) with  $NEI'$  decomposed into forcing and feedbacks,

$$0 = \frac{d}{dy} \left( D \frac{dh'}{dy} \right) + \sum_i c_i T' + S'(1-\alpha) \quad (3)$$

where each  $c_i$  is the (potentially spatially varying) classic feedback parameter for feedbacks linearized in terms of the surface temperature  $T'$ , and we use Cartesian coordinates in the meridional distance  $y$  for simplicity. Roe et al. [31] used the same linearization of feedbacks in a MEBM. We express  $h'$  in terms of  $T'$  using the Clausius-Clapeyron relation, so that (3) becomes

$$\frac{d}{dy} \left( \kappa \frac{dT'}{dy} \right) + CT' = -S'(1-\alpha) \quad (4)$$

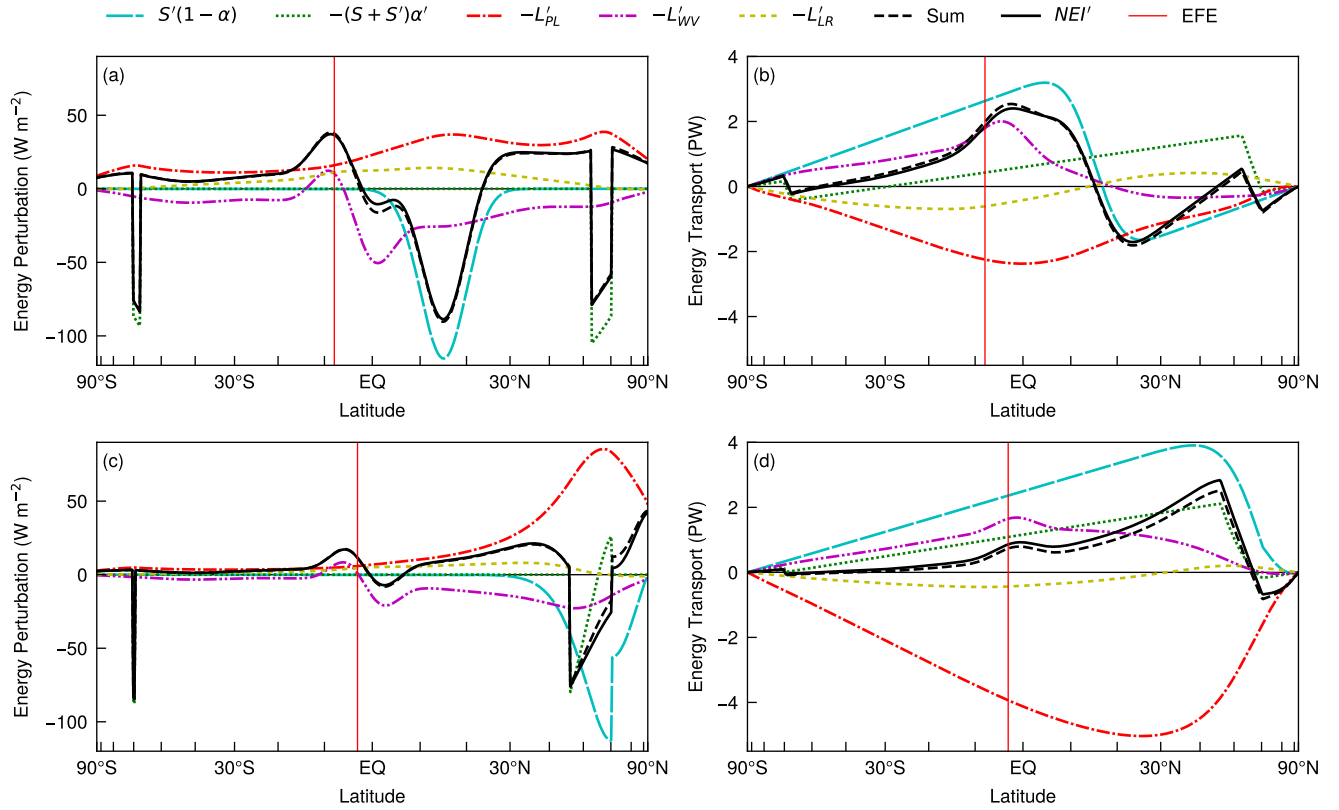


Figure 4: Feedback components of longwave and shortwave responses to a forcing and their associated transports. Panels **(a)** and **(b)** show the anomalies and transports, respectively, for a tropical  $M = 15$  forcing. The same is shown in **(c)** and **(d)** for an extratropical  $M = 15$  forcing. “Sum” refers to the sum of PL, WV, and LR components and should match  $NEI'$  to first order.

where

$$\kappa = D \cdot \left( c_p + L_v r \frac{\partial q^*}{\partial T} \right) \quad (5)$$

and  $C = \sum_i c_i$ . Here,  $c_p$  is the specific heat of air at constant pressure,  $L_v$  is the latent heat of vaporization,  $r$  is relative humidity, and  $q^*$  is the saturation specific humidity. The Clausius-Clapeyron relation provides the value of  $\partial_T q^*$ , which increases with basic-state temperature at a rate sufficiently rapid to cause  $\kappa$  to increase by a factor of 3 between the poles and the equator in our MEBM (not shown).

When expressed in terms of temperature, the perturbation form of our MEBM thus has an apparent diffusivity  $\kappa$  that is larger at warmer basic state temperatures. This is a known feature of EBMs that impose a uniform diffusivity on moist static energy [e.g. 32] and has been shown to be relevant to the polar amplification of warming that occurs in response to a globally uniform positive radiative forcing [e.g. 31, 30]. Here we recognize that a smaller temperature diffusivity will result in a  $T'$  response that is more spatially confined and of higher amplitude, thus yielding a PL feedback that is more equatorially asymmetric for extratropical forcings. Specifically, equation (4) is a simple ODE with Green’s function

$$G(y, z) = -\frac{1}{2\kappa\gamma} e^{-\gamma|y-z|}$$

where  $\gamma = \sqrt{-C/\kappa}$  is real since  $C < 0$ . Therefore, the solution to (4) is  $T' = G * (-R_f)$  where  $*$  is the convolution over all space. For localized forcings such as those used here, this convolution will have maximum magnitude at the center of the forcing and  $T'$  will look like a smoothed version of the forcing (we do not plot a full analytical solution here, which would require treatment of the meridional structure in  $C$  and

$\kappa$ , as well as consideration of the spherical domain and polar boundary conditions). The amplitude and meridional width of  $T'$  are controlled by the magnitudes of  $\kappa$  and  $C$ ; for fixed  $C$ , increasing  $\kappa$  causes  $G$  to spread out, reducing the amplitude and increasing the meridional width of  $T'$ . Since the PL feedback is negative and since all feedbacks produce larger EFE shifts when their contribution to  $NEI'$  is more localized on one side of the equator, we expect the PL feedback to have a greater damping effect on EFE shifts for smaller values of  $\kappa$ , which in turn are found in colder (i.e. extratropical) parts of the planet. This is consistent with the PL feedback on extratropical forcings having a contribution to  $NEI'$  that is more localized (Fig. 4c) and a flux form with greater magnitude on the equator (Fig. 4d).

This enhanced localization of  $T'$  anomalies and the associated larger effect of the PL feedback on EFE shifts occurs in the absence of any rotational confinement of temperature anomalies in our diffusive MEBM. This contrasts with arguments of Seo et al. [23], who suggested that a larger Coriolis parameter at higher latitudes would create larger  $T'$  gradients and thus contribute to the PL feedback being more confined for extratropical forcings. Furthermore, energy diffusion in the absence of feedbacks cannot explain the smaller sensitivity to extratropical forcings in our MEBM, because the flux form of the forcing,  $\mathcal{T}_{S'(1-\alpha)}$ , represents the compensating energy transport that would occur in the absence of all feedbacks, and this has nearly the same equatorial value for tropical and extratropical forcings (compare cyan lines in Fig. 4b, d). Finally, we note that the linearized Planck feedback parameter  $c_{PL}$  (which is distinct from the EFE sensitivity parameter  $\lambda_{PL}$  presented in Fig. 3) is expected to be smaller at colder basic-state temperatures, as can be seen by linearizing the Stefan-Boltzmann law:  $L' \simeq 4\sigma T^3 T'$ . This temperature-dependence

of  $c_{PL}$  should compensate for some of the decrease in  $\kappa$  that occurs in the colder extratropics.

## A relative humidity feedback

Why is the MEBM less sensitive to tropical forcings than both GCMs, and why does the MEBM respond more linearly to both tropical and extratropical forcings than the GCMs? Examination of the humidity distribution in CESM2 provides a possible answer to these questions: the forcings produce large relative humidity (RH) changes that go beyond a simple meridional shift of the basic state distribution (compare solid and dash-dotted lines in Fig. 5a). In our default MEBM, the meridional distribution of RH was constructed to change only by shifting meridionally with the EFE, which would naively seem to capture what C18 describe as the main cause of the WV feedback on EFE shifts: “the shift of the water vapor-rich ITCZ into the hemisphere with greater net column heating.” However, as the ITCZ shifts southward in CESM2, the low-RH region north of the ITCZ broadens and dries while the low-RH region south of the ITCZ contracts and moistens.

To investigate how important these changes in RH of the dry zones are for the WV feedback, we parameterize them in the MEBM using an empirical fit to CESM2. For all  $M$  values, we calculate the best linear fit between the EFE latitude and the 600 hPa zonal mean RH minimum in each hemisphere (Fig. 5b). Then, instead of simply shifting the basic-state RH distribution in the MEBM (that basic state RH is shown by the dashed line in Fig. 5a), we additionally specify the northern and southern RH minima to vary linearly with EFE latitude. For the example in which  $\phi_E = 10.7^\circ\text{S}$  (corresponding to an  $M = 18$  tropical forcing for CESM2), this parameterization produces an equatorial asymmetry in RH in the MEBM that is qualitatively similar to but modestly stronger than that in CESM2 (compare dotted and dash-dotted lines in Fig. 5a). The EFE sensitivity in the MEBM nearly doubles as a result of including this RH parameterization (upward-pointing triangles in Fig. 2), becoming even stronger than that seen in the GCMs (note that these new MEBM simulations also include all other feedbacks). The WV feedback on EFE shifts thus cannot be represented using traditional theory of fixed-RH feedbacks employed for global mean temperature changes, nor can it be thought of as a simple translation of the high-RH region of the ITCZ into the anomalously heated hemisphere.

## Sensitivity to the structure of diffusivity

We saw above that the meridional structure of the effective temperature diffusivity can alter the strength of the PL feedback on EFE latitude. Might meridional variations in  $D$ , the moist static energy diffusivity in equation (1), be important? Previous studies have shown that if  $D$  is diagnosed from idealized GCMs, it is lower in the tropics than in the midlatitudes [e.g. 33, 34]. Following those studies, we calculate  $D(x)$  for our CESM2 simulations as the coefficient that linearly relates the zonal mean, vertically integrated atmospheric energy flux to the zonal mean  $h$ ; it varies by nearly an order of magnitude, with maxima in the midlatitudes and minima at both the equator and poles (Fig. 6a, solid line). The uniform value of  $D$  used in our default MEBM [black line, taken from 29] approximates the global mean diffusivity in CESM2, but differs from the local  $D$  by a factor of five at the equator.

Imposing the distribution of  $D(x)$  from CESM2 in our MEBM produces a large weakening of EFE sensitivity to both tropical and extratropical forcings (compare orange crosses to black circles in Figs. 6b, c). For tropical forcings, this reduction in sensitivity is as strong as that obtained by removing the WV feedback.

Motivated by this result, we hypothesize that the diffusivity near the EFE has the largest impact on the EFE sensitivity to a forcing. To test this hypothesis, we perform two additional sets of MEBM simulations with diffusivities  $D_1(x)$  and

$D_2(x)$ , which are square waves with maxima in the tropics and extratropics, respectively, both having the same global mean as the uniform  $D$  imposed in our default MEBM (Fig. 6a, dashed and dash-dotted lines). For both tropical and extratropical forcings, simulations using  $D_1$  have sensitivities at least as large as those of the constant  $D$  simulations, while simulations using  $D_2$  have smaller sensitivities (hexagons and diamonds in Figs. 6b, c). The MEBM simulations using  $D_2$  have sensitivities nearly the same as the MEBM simulations using the exact  $D(x)$  profile diagnosed from CESM2, except for the much larger EFE shift produced for the tropical  $M = 18$  forcing; that particular simulation is so cold that sea ice expands to  $10^\circ\text{N}$ , causing the AL feedback to dominate.

The meridional structure of diffusivity can thus be just as important as climate feedbacks in setting the EFE response to a forcing. A larger near-equatorial diffusivity could produce a larger EFE sensitivity in at least two ways: by increasing the anomaly of cross-equatorial transport induced by a given forcing, or by flattening the basic-state cross-equatorial energy transport and thus increasing the EFE shift that occurs for a given anomaly of cross-equatorial transport [e.g. 10]. The former effect dominates in our MEBM (not shown), seemingly because a larger near-equatorial diffusivity allows  $h'$  and gradients in  $h'$  (which are proportional to the energy transport) to spread to the equator rather than being confined to off-equatorial regions.

## DISCUSSION

Here we used an MEBM to investigate the influences of climate feedbacks and the eddy-diffusivity of moist energy on the sensitivity of the ITCZ to remote inputs of energy through the top and bottom of the atmosphere. The water vapor, lapse rate, Planck, and albedo feedbacks on ITCZ shifts all have the same sign as they do in the context of the global mean surface temperature response to a global mean radiative forcing, but the magnitude of these feedbacks on ITCZ shifts depends on whether the forcing is localized in the tropics or extratropics. The albedo feedback is positive and local: it is of moderate strength for extratropical forcings but is negligible for tropical forcings. The lapse rate feedback is negative and is the weakest of all the feedbacks examined here, except for the albedo feedback on tropical forcings.

Two of our most novel results concern the Planck and water vapor feedbacks. As shown in previous work [e.g. 32, 30], for a uniform moist energy diffusivity, the effective temperature diffusivity is smaller in colder parts of the planet. This makes the local temperature response, and thus the Planck feedback, larger for extratropical forcings than for tropical ones. In our cloud-free MEBM, this is the primary reason why the ITCZ shift is larger for tropical forcings. The Planck feedback is opposed by the water vapor feedback, which in the context of the global mean climate sensitivity is known to result from the change in water vapor mixing ratio that occurs at fixed relative humidity [e.g. 35]. Here we find, through comparison of the MEBM with a GCM, that the water vapor feedback on ITCZ shifts is instead dominated by an increase in the cross-equatorial asymmetry of the relative humidity of the subtropical dry zones, with additional contributions by the changes in water vapor that occur at fixed-relative humidity and as the high-relative humidity ITCZ shifts meridionally.

Finally, we showed that ITCZ shifts are highly sensitive to the meridional structure of the diffusivity used for moist energy. The ITCZ shift increases with the magnitude of the tropical diffusivity, even when the global mean diffusivity is fixed. Given that the moist energy diffusivity of GCMs is known to be state-dependent [e.g. 34], it seems plausible that diffusivity changes will have a large effect on ITCZ shifts. The diffusivity that acts on anomalies induced by a forcing may even be different from the diffusivity that sets the basic state.

The results from our MEBM show that a given sensitivity



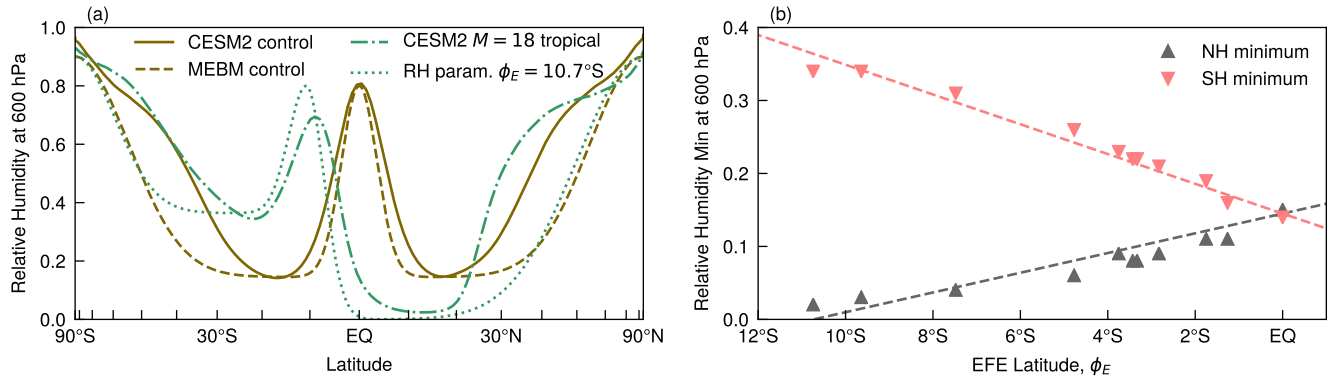


Figure 5: Parameterization of an RH feedback from CESM2. **(a)** Latitudinal profile of RH at 600 mb for CESM2 and the MEBM with RH feedback. The CESM2  $M = 18$  tropical forcing simulation (dash-dotted line) has an EFE of  $\phi_E = 10.7^\circ\text{S}$ , which we input to our RH feedback parameterization to obtain a profile for the MEBM corresponding to the dotted line. **(b)** Linear trends of RH minima at 600 mb versus EFE shift for CESM2 simulations. The southern hemisphere (SH) minimum follows a slope of  $-0.0204$  with  $r^2 = 0.97$ , and the northern hemisphere (NH) minimum slope is  $0.0145$  with  $r^2 = 0.83$ . Both fits are forced through the intercept of  $0.145$  for the control simulation (where  $\phi_E = \text{EQ}$ ).

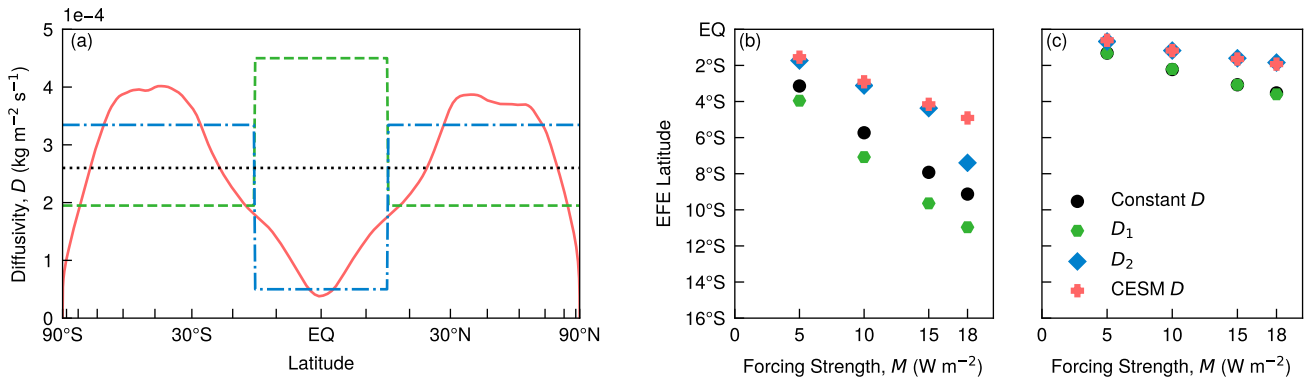


Figure 6: MEBM EFE sensitivity for different prescribed temperature diffusivities,  $D(x)$ . **(a)** Meridional structure of the four diffusivities used. The constant  $D$  (dotted) is the same as in Hwang and Frierson [29] and Roe et al. [31], using the units of the latter. Also shown are  $D$  diagnosed from CESM2 (solid),  $D_1$  (dashed), and  $D_2$  (dash-dotted). EFE shifts using these diffusivities are shown in **(b)** for tropical forcings and **(c)** extratropical forcings.

of the ITCZ to remote forcings in a model can result from large, compensating biases in the diffusivity and feedbacks in that model. Our default MEBM may have such compensating biases due to its use of a uniform diffusivity and a weak WV feedback that does not modify the RH of the subtropical dry zones; a more faithful match to the GCMs might be obtained using a diffusivity that is small in the tropics and high in midlatitudes (e.g.  $D_2$ ) together with a strong RH feedback. Similar reasoning might apply when comparing idealized GCMs with the observed atmosphere if, for example, the effective diffusivity is dominated by transient eddies in the former but by stationary eddies in the latter. Our results thus show that quantitatively understanding radiative feedbacks may make it possible to truly predict, rather than merely diagnose, the ITCZ response to a forcing, but that such an effort may also require knowledge of the effective diffusivity that acts on anomalies of both moist energy and temperature.

## METHODS

### Energy Balance Model

The MEBM, as given in equation (1), is represented using a second-order finite difference scheme solved with the multigrid method. As in previous studies [29, 32, 31, 36], this equation represents the balance of the moist energy of a single, bulk atmospheric layer without an explicit ocean or land surface. For simulations with uniform diffusivity, we use  $D = 2.6 \times 10^{-4} \text{ kg m}^{-2} \text{ s}^{-1}$ . We use a hierarchy of six grids with uniform grid spacings of  $\Delta x = 1/512, 1/256, 1/128, 1/64, 1/32$ , and  $1/16$ , using injection to go from finer to coarser grids and bilinear interpolation to go from coarser to finer grids. The model iterates through  $V$ -cycles until changes in  $h$  are sufficiently small.

The function  $NEI$  can be arbitrarily complex in our MEBM, and most of our simulations use a sophisticated longwave radiative transfer scheme while neglecting atmospheric shortwave absorption. This allows for study of a simple ice-albedo feedback and several longwave radiative feedbacks. We omit the radiative effects of clouds, in part because the cloud-radiative feedback is one of the better studied feedbacks on

ITCZ shifts [e.g. 8, 24, 37]. In particular, we use

$$S(x) = \frac{S_0}{\pi} \sqrt{1 - x^2}, \quad (6)$$

where  $S_0 = 1365 \text{ W m}^{-2}$ , as a simplified insolation, following C18. There are no diurnal or seasonal cycles in the MEBM. The effect of surface ice cover (e.g. ocean sea ice) on albedo is parameterized by

$$\alpha(x) = \begin{cases} 0.2, & T(x) < -10^\circ\text{C} \\ 0.6, & T(x) > -10^\circ\text{C} \end{cases} \quad (7)$$

as in North [38]. We repeated many of our simulations using a more complicated shortwave radiative transfer scheme similar to that described below for longwave, but like C18, we did not see appreciable differences in the EFE shifts. We thus opted for the simpler treatment of shortwave, described above, for simplicity.

To represent  $L$ , we use the Rapid Radiative Transfer Model for GCMs (RRTMG) provided in the climt (Climate Modelling and Diagnostics Toolkit) software package [39], which parameterizes longwave fluxes based on the vertical profiles of temperature and humidity in non-interacting columns with 30 vertical levels. The temperature profile for these columns is taken to be a moist adiabat from surface air temperature  $T$ , which in turn is obtained from  $h$  by assuming a surface air relative humidity (RH) of 0.8.

Above the surface, the RH is prescribed to have idealized vertical and meridional structures drawn from that of GCM aquaplanet simulations (C18 and CESM2, described below). These structures are constructed by first segmenting the atmosphere into a boundary layer (surface to 875 hPa), free troposphere (875 hPa to 200 hPa), tropopause layer (200 hPa and to 100 hPa), and stratosphere (100 hPa and above). In the boundary layer and stratosphere, RH is uniformly set to 0.9 and 0.0, respectively. In the free troposphere and tropopause layers, a meridional structure is set by prescribing RH in terms of Gaussians in  $x$ . For the free troposphere, we use four Gaussians in  $x$  which move together with latitude of the EFE,  $\phi_E$ , allowing for a moist ITCZ-like column at  $\phi_E$  surrounded by dry subtropical zones and humid poles. Specifically, a half-Gaussian in RH is prescribed in each of these subdomains:  $[-90^\circ, -15^\circ + \phi_E]$ ,  $[-15^\circ + \phi_E, \phi_E]$ ,  $(\phi_E, 15^\circ + \phi_E]$ , and  $[15^\circ + \phi_E, 90^\circ]$  with the convention that negative latitudes lie south of the equator. We make the spread of each Gaussian small enough so that the magnitude and derivative of RH are approximately the same at the boundaries of the subdomains. At 600 hPa, this yields the dashed line in Fig. 5a. The RH in the tropopause-layer is prescribed to be a single Gaussian centered at  $\phi_E$  with a spread of  $20^\circ$  latitude, yielding a humid tropics and dry extratropics.

With these choices, the control MEBM has a basic state with surface air temperature varying from 240 K at the poles to 300 K at the equator (Fig. 1b, black line).

## Suppressed-feedback simulations in the MEBM

In additional sets of simulations, we remove individual feedbacks from the forced runs by prescribing the albedo, specific humidity, or lapse rate from the control simulation. For instance, simulations without the AL feedback simply use  $\alpha$  from the equilibrated control simulation; the influence of the AL feedback on the EFE shift is obtained by differencing forced simulations without the AL feedback and forced simulations of the control model. The LR feedback is suppressed by prescribing the vertical lapse rate from the equilibrated control simulation, at each latitude, instead of the moist adiabatic lapse rate corresponding to the temperature achieved in the forced simulation. We suppress the WV feedback by using saturation specific humidity,  $q^*$ , and relative humidity distribution

from the control simulation. One might try to suppress the PL feedback by using surface temperatures from the control simulation, but a steady state would then not be achieved since our forcings have a non-zero global mean. So we instead run a simulation with no WV, LR, or AL feedbacks and another simulation with all feedbacks turned off (by removing the AL feedback and setting  $L' = -M$ ) in order to determine, by difference, the effect of the PL feedback on  $\phi_E$  (see Fig. 3). Essentially, this allows the global mean PL feedback to balance the global mean forcing while removing the effect of the PL feedback on horizontal gradients of the net energy input. We additionally diagnose the PL feedback from an offline OLR calculation with a model state having the forced simulation's water vapor profile and lapse rates, but surface temperatures matching those of the control simulation (see Fig. 4).

## General Circulation Model

We repeated the basic state simulation and the series of forced simulations (described above), with all longwave feedbacks active, using the atmospheric component of the Community Earth System Model version 2.1.0 (CESM2), a three-dimensional GCM with sophisticated parameterizations of moist convection, radiation, turbulence, and other subgrid-scale processes [40, 41]. Specifically, we use the aquaplanet version of the model from the ‘‘Simpler Models’’ initiative [42], with a slab ocean mixed layer and the  $1^\circ \times 1^\circ$  finite-volume dynamical core of the Community Atmosphere Model version 6 (CAM6). We employ a 1-meter deep ocean mixed layer. These simulations are meant to replicate some of the results of C18, but in a different GCM and one that allows us access to detailed model output; we thus follow C18 and eliminate all cloud-radiative interactions but prescribe a globally uniform surface albedo of 0.2725. There is no sea ice and thus no ice-albedo feedback, so the lower ice-free albedo specified in the MEBM (7) allows the two models to achieve a similar global mean temperature. We impose the same insolation as in the MEBM, although the GCM has an equinoctial diurnal cycle with the instantaneous insolation prescribed to give the time-mean distribution specified in (6).

## DATA AVAILABILITY

Output from the simulations is available upon request.

## CODE AVAILABILITY

The MEBM code is freely available at <https://github.com/henrygrantpeterson/mebm>. Contact H.G.P. for questions regarding the code.

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## AUTHOR CONTRIBUTIONS

W.R.B. conceived the project, but both authors designed the study, the modelling strategy, and the analyses. H.G.P. devel-



oped the MEBM. W.R.B. ran CESM2. Both authors wrote the manuscript.

## ADDITIONAL INFORMATION

**Competing interests:** The authors declare no competing interests.

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