

Symmetric and antisymmetric components of polar-amplified warming

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ABSTRACT

We document an intriguing property of zonal-mean surface warming in 3,000-year general circulation model (GCM) simulations under instantaneous 2, 4, 8, or $16 \times \text{CO}_2$ increases: across CO₂ magnitudes and time periods from the first decade to the thirtieth century, the hemispherically symmetric, global-mean-normalized response is quasi-invariant. For a given CO₂ value, the hemispherically antisymmetric component evolves in time from initially weak Southern Ocean warming and strong Arctic warming to comparably polar-amplified warming in the two hemispheres. For a given time period, the mean-normalized, full response (i.e. including both symmetric and antisymmetric components) is remarkably consistent across CO₂ values. Simulations in a moist energy balance model (MEBM) with radiative feedbacks diagnosed from the $4 \times \text{CO}_2$ GCM simulation at different timescales reproduce this quasi-invariant mean-normalized symmetric warming component. This also holds in complementary MEBM simulations with symmetrized inputs, indicating little role for covarying antisymmetric components of the radiative feedback parameter and temperature response. In 800-year GCM simulations wherein CO₂ is unchanged but cloud albedo is artificially modified in different latitude bands, the mean-normalized, symmetric temperature response is again nearly time-invariant, but its meridional structure is influenced by the meridional structure of the imposed albedo perturbations. The non-normalized temperature response at high latitudes, however, depends more on the value of forcing imposed locally than on its meridional structure.

1. Introduction

Climatologically, annual-mean, zonal-mean surface air temperature is highest at low latitudes and decreases appreciably toward the pole in both hemispheres. This hemispherically symmetric planetary-scale signal is much larger than the deviations therefrom in either hemisphere. Fig. 1 illustrates this via a preindustrial control simulation in a climate model (whose formulation will be detailed further below).

The most fundamental driver of this symmetry is surely the insolation distribution, which, annually averaged, is symmetric in latitude about the equator, decreasing from its equatorial maximum of $\sim 400 \text{ W m}^{-2}$ to $\sim 180 \text{ W m}^{-2}$

at either pole. And longwave trapping by the well-mixed greenhouse gases including CO₂ is roughly (though not exactly, c.f. Merlis 2015; Huang et al. 2017) homogeneous and therefore symmetric. Earth's ocean basins, land masses, orography, vegetation cover, sea-ice, and water vapor fields are all zonally and hemispherically heterogeneous — but evidently not enough to yield predominantly hemispherically antisymmetric climatological surface air temperatures. In the bulk sense of Northern Hemisphere vs. Southern Hemisphere averages, a remarkable symmetry in the top-of-atmosphere (TOA) albedo has been documented in observations and climate models, both in the climatology (Voigt et al. 2013; Stephens et al. 2015) and under hemispherically antisymmetric external forcing (Voigt et al. 2014).

Must externally forced surface air temperature patterns be nearly hemispherically symmetric as they are clima-

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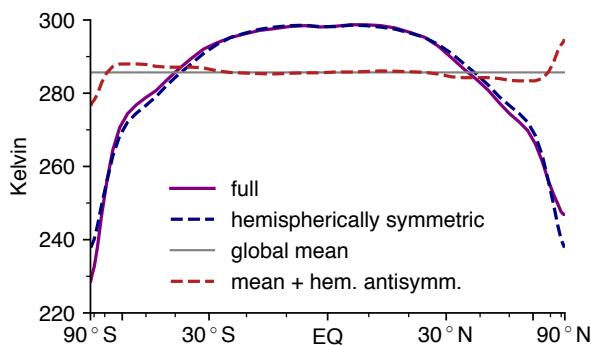


FIG. 1. Climatological annual-mean, zonal-mean surface air temperature in the preindustrial control simulation, the hemispherically symmetric component thereof, the global-mean value, and the global-mean plus the hemispherically antisymmetric component, as indicated in the legend, all in units Kelvin. Horizontal axis spacing in this and all subsequent plots is proportional to sine of latitude and thus surface area.

tologically? Even for a nearly uniform forcing such as increased CO₂, there is no obvious physical mechanism that would enforce this *a priori*. Nevertheless, in this manuscript we document a remarkable consistency in the hemispherically symmetric component of CO₂-forced warming in a climate model across timescales and, normalized by its global-mean value, across forcing magnitudes as well.

Empirically, surface temperature responses to external forcings are almost universally polar-amplified, with larger temperature changes at high than low latitudes in both hemispheres. Polar amplification emerges in general circulation models (GCMs), simple diffusive energy balance models (EBMs), and the real world as reflected in paleoclimate proxy records of past warm and cool states. But it also includes an appreciable antisymmetric signature: in the initial century following a CO₂ increase, mean downwelling of surface water by prevailing Southern Ocean currents inhibits local surface warming (e.g. Armour et al. 2013; Marshall et al. 2015), which combined with weakly negative radiative feedbacks in northern high latitudes results in a predominantly asymmetric, Arctic-amplified transient response (e.g. Stuecker et al. 2018).

Most GCM simulations under increased CO₂ do not extend far beyond this first century due to computational costs (e.g. Held and Soden 2006; Zelinka and Hartmann 2012; Stuecker et al. 2018). But in longer running GCM integrations the high latitudes continue to warm relative to the Tropics (in absolute terms) for centuries and millennia (e.g. Armour et al. 2013; Rugenstein et al. 2019). This multi-millennial equilibrium response is sometimes approximated from simulations with the dynamical ocean replaced by a rapidly equilibrating motionless slab (e.g. Held and Soden 2006; Armour et al. 2013). This negates any role for changes in deep-ocean circulation,

which can be non-monotonic and unfold on a millennial timescale (Jansen et al. 2018). For example, in a 3,000-year GCM simulation with the meridional cloud albedo distribution perturbed so as to generate a surface climate resembling the early Pliocene (~4 Ma, e.g. Burke et al. 2018), a Pacific Meridional Overturning Circulation (PMOC) emerges after ~1,500 years, increasing the heat convergence into the northern hemisphere (Burls et al. 2017).

Here, we investigate the meridional pattern of zonal-mean surface air temperature from decadal to millennial timescales in 3,000-year GCM simulations under CO₂ increases from 2 to 16× preindustrial, in simulations performed with a simple moist energy balance model meant to crudely approximate the GCM CO₂ simulations, and on decadal to multi-centennial timescales in perturbed-cloud-albedo GCM simulations originally presented by Burls and Fedorov (2014b) from which the aforementioned Pliocene-like simulation was selected (Burls and Fedorov 2014a; Fedorov et al. 2015). Section 2 describes the GCM, the perturbed-CO₂ simulations performed, and other methodological choices. Section 3 describes results from the perturbed-CO₂ GCM simulations. Section 4 describes results from the moist energy balance model simulations. Section 5 describes the perturbed-cloud-albedo GCM simulations and results therefrom. Section 6 concludes with summary and discussion.

2. Methods

We analyze simulations performed in the National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM), version 1.0.4, in its low-resolution configuration (Shields et al. 2012). It consists of the Community Atmosphere Model, version 4 (CAM4) with its spectral dynamical core truncated at T31 resolution (~3.75° × 3.75°) and with 26 vertical levels coupled to the Parallel Ocean Program version 2 (POP2) with ~3° horizontal resolution and 60 vertical levels. We refer the interested reader to Shields et al. (2012) for additional details.

A standard preindustrial control simulation spans 3,000 years. The perturbation simulations comprise standard instantaneous CO₂ increase simulations of 2, 4, 8, and 16× preindustrial concentrations, each run for 3,000 years, along with 800-yr simulations in which cloud albedo is modified that are detailed in Section 5.

We analyze temporal averages over years 1-10 and 21-100 (periods during which both the atmosphere and ocean are rapidly responding), 701-800 (during which the atmosphere is in a nearly statistically steady state but the ocean remains slowly varying), and years 2901-3000 (at which time the deep ocean has nearly equilibrated). These are similar to the four timescales identified by Armour et al. (2013) of the climate response to abrupt CO₂ doubling,

although their “equilibrium” solution was obtained using a slab ocean model rather than running the full coupled model to equilibrium. Model drift in the control simulation is modest relative to the forced temperature responses in the CO₂ and perturbed albedo simulations (over the 3,000 years span of the control simulation, global-mean ocean temperature increases by ~ 0.4 K). Therefore, for convenience responses at all timescales in the perturbation experiments are computed with respect to averages of the control over years 701–800.

We decompose the surface temperature response and select other fields into components that are symmetric and antisymmetric about the equator. Symbolically for a given field χ ,

$$\chi(\varphi) = \chi^s(\varphi) + \chi^a(\varphi), \quad (1)$$

where φ is latitude,

$$\chi^s \equiv \frac{\chi(\varphi) + \chi(-\varphi)}{2} \quad (2)$$

is the symmetric component, and

$$\chi^a \equiv \frac{\chi(\varphi) - \chi(-\varphi)}{2} \quad (3)$$

is the antisymmetric component. The symmetric component is an even function of latitude whose global-mean value (which we denote by overbars) is identical to the full field’s global-mean [$\bar{\chi}^s(\varphi) = \bar{\chi}^s(-\varphi)$ and $\bar{\chi}^s = \bar{\chi}$], and the antisymmetric component is an odd function of latitude whose global mean is zero [$\bar{\chi}^a(\varphi) = -\bar{\chi}^a(-\varphi)$ and $\bar{\chi}^a = 0$].

We do not perform a formal radiative feedback analysis, whose technical interpretation can be subtle with respect to mechanisms of polar amplification, perturbed energy transports, and meridional temperature response patterns more generally (e.g. Merlis 2014; Russotto and Biasutti 2020).

3. Results from 2–16×CO₂ simulations

Fig. 2 shows zonal-mean surface air temperature change averaged over years 701–800 in the 2, 4, 8, and 16×CO₂ simulations. Surface warming occurs at all latitudes, is weakest and relatively flat at low latitudes, and increases nearly monotonically moving from low to high latitudes (though in the Southern Hemisphere peaking at a local maximum ranging from $\sim 65^\circ\text{S}$ for 2×CO₂ to $\sim 80^\circ\text{S}$ for 16×CO₂). As a function of the imposed CO₂ increases, the equator-to-pole temperature drop decreases in both hemispheres: going from 2 to 16×CO₂, low-latitude warming ranges from ~ 2 to ~ 11 K, peak SH high-latitude warming from ~ 6 to ~ 25 K, and peak NH high-latitude warming (at the North Pole) from ~ 7 to ~ 34 K. Peak warming in the Arctic relative to that in the Antarctic also increases with CO₂ magnitude.

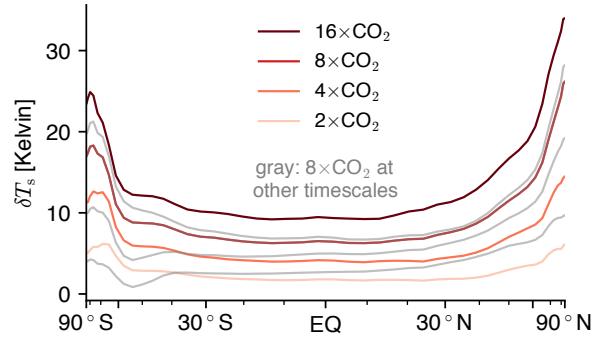


FIG. 2. In red shades according to the legend, annual-mean, zonal-mean surface air temperature response in instantaneous 2, 4, 8, and 16×CO₂ simulations averaged over years 701–800 compared to a preindustrial control. Thin, gray curves show the annual-mean, zonal-mean surface air temperature response in the 8×CO₂ case averaged over the three other time periods considered in detail in this manuscript, from bottom to top years 1–10, 21–100, and 2901–3000.

Fig. 2 also shows the zonal-mean surface air temperature responses for the 8×CO₂ simulation at the other three selected timescales. In the first decade under 8×CO₂, warming in the Southern Hemisphere is closest to the 2×CO₂ case but in the Northern Hemisphere is closest to the 4×CO₂ case, and its dip over the Southern Ocean due to the drawing down of heat from the surface by climatological downwelling is conspicuously absent in the 2×CO₂, year 701–800 response. That dip and hemispheric asymmetry persist over years 21–100, when the 8×CO₂ response falls closest to the 4×CO₂ response of years 701–800 in the Southern Hemisphere and essentially halfway between the 4 and 8×CO₂ year-701–800 responses over most of the Northern Hemisphere. After multiple millennia when the deep ocean has nearly equilibrated (years 2901–3000), the 8×CO₂ response has warmed only slightly further in the Tropics and Northern Hemisphere but more appreciably, ~ 3 K, in the southern high latitudes.

We may summarize this comparison of responses across forcing magnitudes vs. across timescales as follows, taking the liberty to assume that the behavior spanning the gaps between the selected forcing magnitudes and timescales is sufficiently linear: comparing pairs in forcing magnitude and averaging period in which the global-mean surface air temperature responses are nearly the same, those that occur earlier, during the rapid transient response stage, are more influenced by the hemispherically antisymmetric processes that retard Southern Ocean warming and amplify Arctic warming. Those in later centuries reflect a more hemispherically symmetric response, with Antarctic warming catching up to (or exceeding) Arctic warming over multiple millennia.

Fig. 3 repeats Fig. 2 but including all four time periods considered and with each shown both in raw terms and normalized by the global-mean surface air temperature response for that particular CO₂ value and timescale. With this normalization, the full responses are quite similar across the simulations throughout the low latitudes ($\sim 40^{\circ}\text{S}$ - 20°N) at all timescales and up to $\sim 50^{\circ}\text{S}$ and to the North Pole at subsequent timescales. At each timescale, the 2 \times CO₂ case is least like the others, being well separated in the northern mid-latitudes in the first decade and at southern high latitudes in all periods (though years 701-800 least of all). The relative warming of the Arctic vs. Antarctic at (near) equilibrium varies steadily with the CO₂ value: under 2 \times CO₂, the Arctic-Antarctic warming difference effectively vanishes by years 701-800 and reverses at equilibrium, under 4 \times CO₂ it stays near zero at equilibrium, and under 8 and 16 \times CO₂ it retains its original sign throughout.

Each thin curve underlain in Fig. 3 is the hemispherically symmetric component of the corresponding thicker curve. For all forcing magnitudes, the initial hemispheric antisymmetry of the response steadily weakens as Southern Ocean (and to a lesser extent Antarctic) warming catches up with the rapid Arctic warming. At the millennial timescale when the full coupled system has very nearly reached its new equilibrium, the full and symmetric responses lie nearly on top of one another. Moreover, normalized by the global-mean warming this symmetric component appears to change little across both forcing magnitudes and timescales.

Fig. 4(a) emphasizes this by showing on one panel all sixteen symmetrized, mean-normalized surface warming patterns, one for each CO₂ value and averaging period. Apart from the first decade of the weakest perturbation, the mean-normalized symmetric components collapse onto nearly a single curve, with warming less than the global-mean equatorward of $\sim 45^{\circ}\text{S/N}$ and greater than the global-mean poleward thereof. To first order, the quasi-universal curve could be approximated by three linear segments: a nearly flat, Tropical segment ($\sim 0.7\text{ K/K}$ for $\sim 25^{\circ}\text{S}$ - 25°N), a moderately sloped, mid-latitude segment (from ~ 0.7 at $\sim 25^{\circ}\text{S/N}$ to $\sim 1.2\text{ K/K}$ at $\sim 55^{\circ}\text{S/N}$), and a steeply sloped, high-latitude segment (from ~ 1.2 at $\sim 55^{\circ}\text{S/N}$ to $\sim 2.0 - 2.4\text{ K/K}$ at either pole $\sim 55^{\circ}\text{S/N}$). For all CO₂ values the mean-normalized signal becomes more polar amplified in time, with low-latitude values decreasing and high-latitude values increasing. Conversely, equilibrium polar amplification steadily decreases with CO₂ magnitude (except very near the poles).

Fig. 4(b) repeats Fig. 4(a) but for the antisymmetric, mean-normalized surface warming components (note magnified vertical axis spacing in the panel b). These do not collapse onto a quasi-universal curve as do their symmetric counterparts. But their behavior is still relatively straightforward, reflecting the gradual catching-up

of Antarctic and Southern Ocean warming with (and for 2 and 4 \times CO₂, surpassing) the initially rapid Arctic warming. Particularly in the initial, transient century, the curves group together more by timescale than by CO₂ value (in Fig. 4(b), compare curves with the same line markings to those with the same color). This would seem to reflect the timescales intrinsic to the physical processes underlying these responses — no matter how large a TOA forcing over the Southern Ocean, prevailing downwelling of surface water inhibits initial local surface warming, while the deep ocean equilibration that acts to homogenize subsurface warming between the hemispheres take millennia.

Fig. 5 shows maps of the mean-normalized, annual-mean surface air temperature response in each CO₂ simulation and for each time period. The enhanced warming at equilibrium in the southern high latitudes under 2 \times CO₂ is centered just off the Antarctic coast, primarily over the Weddell and Ross Seas. Nevertheless, the 2 \times CO₂ simulation is the only one in which southern hemisphere sea ice does not disappear entirely; under 16 \times CO₂ it is nearly gone by years 21-100, under 8 \times CO₂ by years 701-800, and under 4 \times CO₂ by years 2901-3000 (not shown).

We deem the near-collapse of the mean-normalized, symmetric components onto one pattern worth further consideration. A strong similarity between the spatial patterns of the sea surface temperature (SST) response in the Indo-Pacific ocean in experiments with different CO₂ forcings has been noted, but not fully explained, by Heede et al. (2020). And some collapse of zonal-mean surface air temperature responses at different timescales when mean-normalized can be inferred in the existing literature — across timescales in one GCM (c.f. Fig. 4a of Armour et al. 2013) and across GCMs at individual timescales (c. f. Fig. 6 of Rugenstein et al. 2019). But to our knowledge this stronger collapse of the hemispherically symmetric component has not been shown. There is no obvious *a priori* reason why it should occur, and we will show in Section 5 that it also holds for the perturbed albedo simulations — though with its precise meridional structure influenced by that of the imposed forcing.

4. Moist Energy Balance Model

To gain some physical insight into the quasi-invariance of the symmetric, mean-normalized warming pattern across CO₂ concentrations and timescales, we examine the solutions of a moist energy balance model (MEBM). MEBMs have been a useful simplified modelling framework to develop theory for the spatial pattern of warming in idealized contexts (Flannery 1984; Rose et al. 2014; Roe et al. 2015; Merlis and Henry 2018; Russotto and Biasutti 2020), to emulate the warming pattern in comprehensive GCMs (Hwang et al. 2011; Bonan et al. 2018), and to investigate changes in atmospheric energy transport

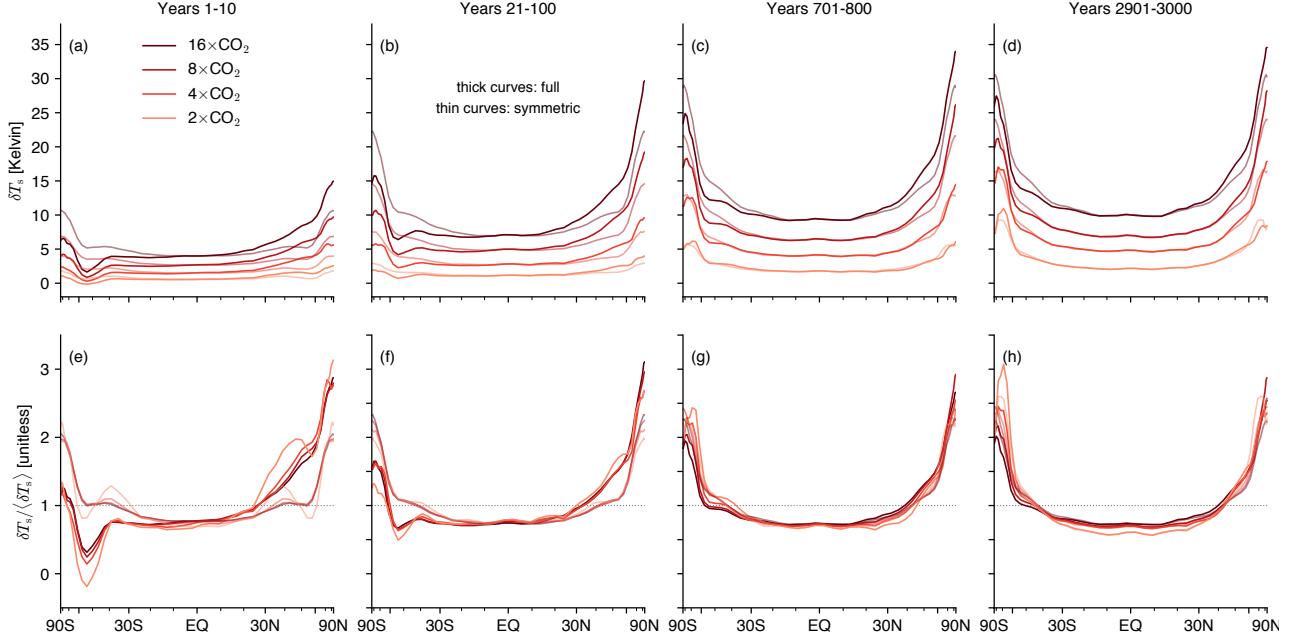


FIG. 3. Annual-mean, zonal-mean surface air temperature response in the 2, 4, 8, and 16 \times CO₂ simulations averaged over (columns, left to right) years 1–10, 21–100, 701–800, and 2901–3000, either (top row, units Kelvin) not normalized or (bottom row, unitless) normalized by the global-mean surface air temperature response in that simulation and time period. Thick curves in each panel are the full fields, and underlain thinner curves are the corresponding hemispherically symmetric components.

(Frierson et al. 2007; Hwang and Frierson 2010; Armour et al. 2019).

a. Model formulation

The MEBM's governing equation is

$$\mathcal{C}\partial_t T = \mathcal{F} + \lambda T + \mathcal{O} + \mathcal{D}\nabla^2 h, \quad (4)$$

where \mathcal{C} is the surface layer heat capacity, T is surface temperature, \mathcal{F} is the imposed radiative forcing, λ is the radiative feedback parameter, \mathcal{O} is the net surface flux (signed positive downward; also known as ocean heat uptake), \mathcal{D} is the diffusivity, and h is surface moist static energy (MSE). All variables in (4) are perturbations about a control climate. In short, the time tendency of the heat content of the surface layer (LHS) is determined by the combined effect of (RHS terms, left to right) the imposed radiative forcing, a radiative restoring term that varies linearly with the surface temperature anomaly, an imposed ocean heat uptake field, and the convergence of the anomalous column-integrated MSE flux, approximated as down-gradient diffusion of surface MSE. We now detail each of the RHS terms.

For the CO₂ radiative forcing, $\mathcal{F}(\phi)$, we use the spatially varying instantaneous forcing of Huang et al. (2016) computed for a doubling of CO₂. Strictly speaking, this will not be identical to the radiative forcing computed

with our particular GCM due to dependencies on the climatology (e.g. Merlis 2015; Huang et al. 2017). But we do not expect such discrepancies to meaningfully impact the results. In order to convert this instantaneous 2 \times CO₂ radiative forcing field into a stratosphere-adjusted 4 \times CO₂ radiative forcing, we double it and add a uniform value of 2.4 W m⁻². This yields a global-mean value of 7.0 W m⁻².

The feedback parameter λ is diagnosed using the radiative forcing field \mathcal{F} just described along with fields taken from the GCM 4 \times CO₂ simulation. Specifically,

$$\lambda(\phi) = -\frac{\mathcal{F}(\phi) - \mathcal{T}}{T_{gcm}(\phi)}, \quad (5)$$

where \mathcal{T} is the anomalous TOA radiative flux in the GCM (signed positive downward), and T_{gcm} is the anomalous surface air temperature in the GCM. The “gcm” subscript is meant to emphasize that this temperature field in the denominator of (5) is that diagnosed from the GCM, not the MEBM's own computed temperature (whereas the temperature field that λ multiplies in (4) is that of the MEBM). One MEBM simulation is performed for each of the four time periods of interest, each with \mathcal{T} and T_{gcm} taken from the GCM 4 \times CO₂ simulation averaged over that time period. Likewise, the prescribed ocean heat uptake field \mathcal{O} is simply diagnosed from the 4 \times CO₂ GCM simulation for the given time period.

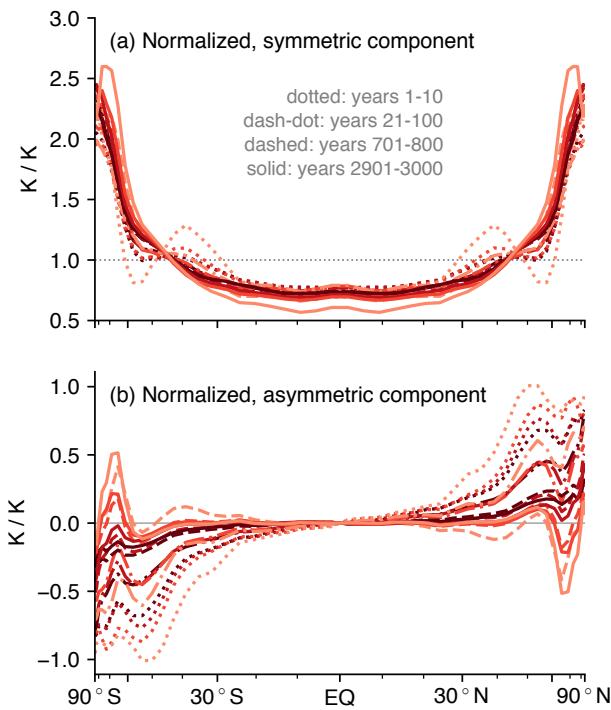


FIG. 4. Annual-mean, zonal-mean surface air temperature response in the 2, 4, 8, and 16 \times CO₂ according to the legend in Fig. 2 averaged within each of the four time periods of focus, with dotted, dash-dotted, dashed, and solid curves corresponding to years 1-10, 21-100, 701-800, and 2901-3000, respectively, normalized by the global-mean response for that forcing and time period, and decomposed into hemispherically (a) symmetric and (b) antisymmetric components. Note that the vertical axis range and spacing differs between the two panels.

Finally, the diffusive approximation to atmospheric energy transport convergence, $\mathcal{D}\nabla^2 h$ is computed as follows. Because all quantities are anomalies, surface MSE is a linearized form $h = T(1 + \mathcal{H}L\partial_T q^*)/c_p$, with relative humidity \mathcal{H} , saturation specific humidity q^* , and latent heat of vaporization L . The partial derivative of the saturation vapor pressure, $\partial_T q^*$, is evaluated using the zonal-mean climatological surface air temperature from the GCM (i.e. surface air temperature averaged over years 701-800 in the preindustrial control simulation). The parameter values for all constant coefficients are standard: $\mathcal{H} = 0.8$, $\mathcal{D} = 0.3 \text{ W m}^{-2} \text{ K}^{-1}$, $c_p = 1004.6 \text{ J kg}^{-1} \text{ K}^{-1}$, and $L = 2.5 \times 10^6 \text{ J kg}^{-1}$.

Across the four MEBM simulations, each of which corresponds to a different time period of the 4 \times CO₂ simulation, the only parameters that differ are λ and \mathcal{O} as just described. The MEBM is integrated in time using a fourth-order Runge-Kutta scheme to equilibrium, while a second-order finite difference scheme is used for the ∇^2 operator (Wagner and Eisenman 2015). There are 60 model grid points evenly spaced in $\sin \varphi = 1/30$ increments, with

gridpoint centers in each hemisphere from $\sin \varphi \approx 0.12$ (corresponding to $\varphi \approx 4.8^\circ$) to $\sin \varphi \approx 0.98$ (corresponding to $\varphi \approx 79.5^\circ$). The GCM fields are spaced evenly in latitude, with 48 points total, spanning in each hemisphere from approximately 1.8 to 87.2° with approximately 3.6° spacing. The GCM fields that are used as inputs to the EBM, therefore, are spectrally transformed at order 20 to the EBM grid.

b. Results

Fig. 6 shows the surface air temperature response at each time period in the GCM 4 \times CO₂ simulation along with that of the corresponding MEBM simulations. In all cases, the MEBM captures the overall warming pattern reasonably well (panel a). The most prominent bias of the MEBM is an insufficiently sharp warming gradient at high latitudes, particularly in the Southern Hemisphere. This is a common feature of MEBMs with a uniform diffusivity (e.g. Bonan et al. 2018). When the results are normalized by their global mean (panel b), these high-latitude deficiencies of the MEBM simulations are again pronounced and particularly in the Southern Hemisphere. But the MEBM captures each global-mean response accurately enough, in addition to the spatial pattern, yielding reasonably accurate mean-normalized fields.

The weaker slope of warming with latitude in the extratropics in the MEBM compared to the GCM projects onto the hemispherically symmetric component of the mean-normalized warming pattern as well (panel c). But otherwise, again the MEBM captures the features of our interest from the GCM, in particular the quasi time-invariance. It also captures the decrease with time at low latitudes and an increase with time at high latitudes, i.e. increasing polar amplification. Finally, the MEBM also captures the general character for each time period of the mean-normalized hemispherically antisymmetric component (panel d), with initial warming of the Arctic relative to the Antarctic, a reduction of this hemispheric difference over subsequent decades and centuries, and after several centuries portions of the Antarctic that have warmed more than the corresponding Arctic latitudes (more so for the MEBM than the GCM).

c. MEBM simulations with hemispherically symmetrized inputs

Because the MEBM simulations just described capture the temporal quasi-invariance of the mean-normalized, hemispherically symmetric warming pattern from the GCM 4 \times CO₂ simulation, we perform an additional set of MEBM to elucidate the underlying physical mechanisms.

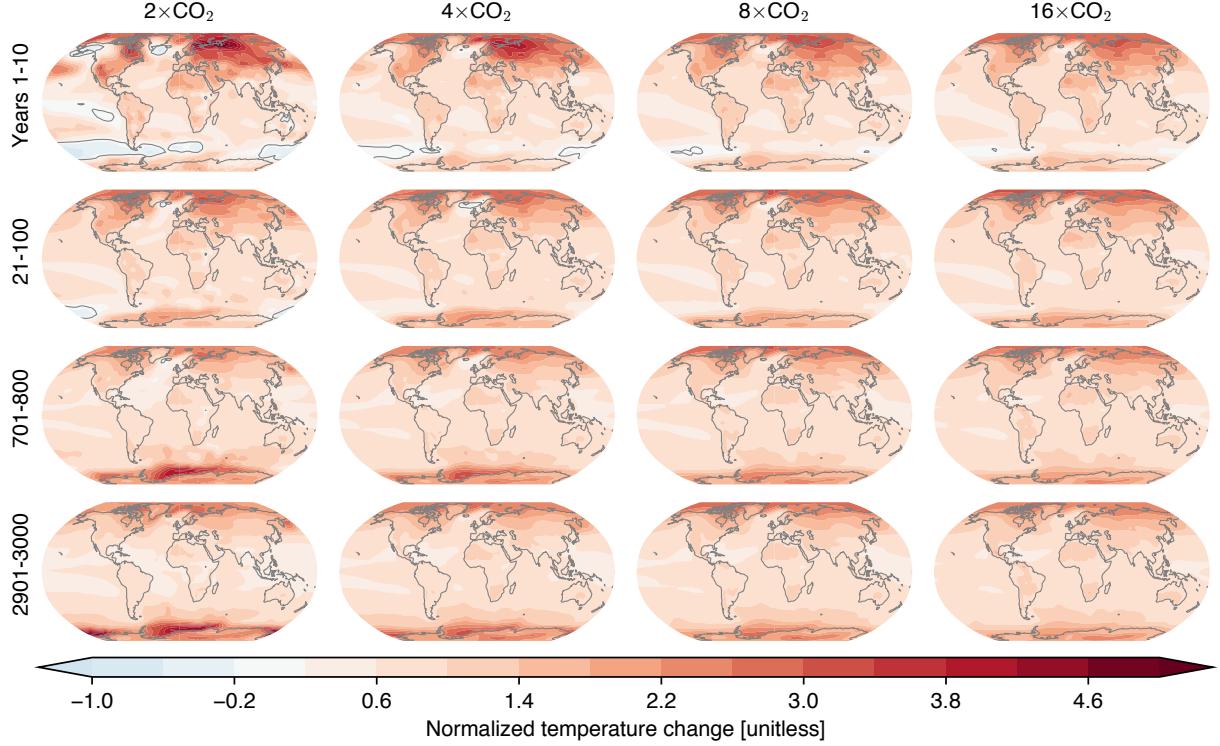


FIG. 5. Global-mean-normalized, annual-mean surface air temperature response in the 2 , 4 , 8 , and $16 \times \text{CO}_2$ GCM simulations according to the colorbar, each shown separately for each of the four time periods, as indicated by the column and row labels. The zero contour is shown as a thin gray contour.

Each MEBM simulation is repeated but with hemispherically symmetrized versions of all of its input fields (radiative forcing, the radiative feedback parameter, the net surface energy flux, and the climatological surface air temperature used to compute $\partial_T q$). In these symmetrized-inputs simulations, there is no possibility of covarying hemispherically antisymmetric components of different terms projecting onto the hemispherically symmetric component of the temperature field.

To illustrate this, Fig. 7 shows the full, symmetric, and antisymmetric components of the radiative feedback parameter diagnosed for each of the four time periods for $4 \times \text{CO}_2$. The full field (panel a) is negative at nearly all latitudes in all time periods, with the smallest negative to slightly positive values occurring in the southern high latitudes. Values at low latitudes are more negative overall, except in the first decade when the global minimum of approximately $-6 \text{ W m}^{-2} \text{ K}^{-1}$ occurs over the Southern Ocean. This is the largest antisymmetric signal (panel c) of any time period or latitude band, with the antisymmetric component otherwise being generally smaller magnitude than the symmetric component (panel b). The symmetric component also reveals that the feedback parameter becomes less negative at most latitudes with time, as has been documented previously (Armour et al. 2013).

Fig. 8 shows the temperature response in the MEBM full simulation, the hemispherically symmetric component thereof, and the temperature response in the symmetrized-inputs MEBM simulation. For each time period, the symmetrized temperature response of the full simulation is nearly identical at all latitudes to the temperature response of the corresponding symmetrized-inputs simulation. The biggest discrepancies (though still modest, $< 1 \text{ K}$ everywhere) are for the first decade, consistent with the antisymmetric components of the input and temperature fields having their largest magnitudes relative to the symmetric component in this time period.

d. Symmetrized GCM fields that determine symmetric input

The results of the symmetrized-inputs simulations suggest that understanding the quasi-invariance of the mean-normalized, symmetric warming component can be understood entirely through hemispherically symmetric processes within the MEBM. Fig. 9 repeats Fig. 7 but for the net surface flux rather than the radiative feedback parameter. The global mean of \mathcal{O} decreases from 3.9 W m^{-2} in the first decade to near zero, -0.4 W m^{-2} , during the final, near-equilibrium period. The meridional slope from

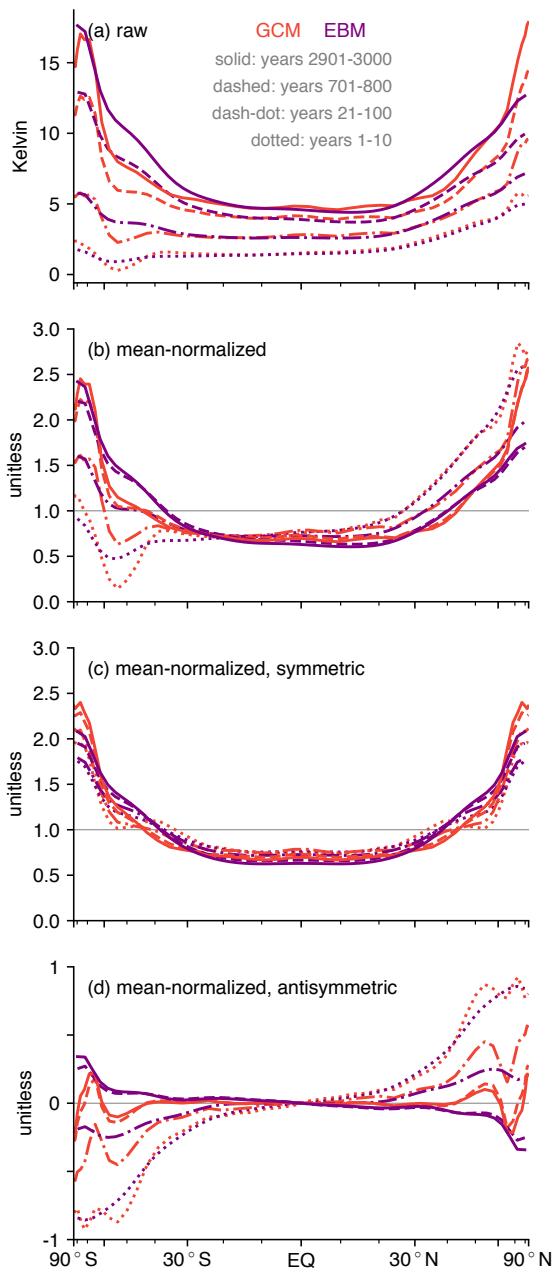


FIG. 6. Surface air temperature response in the GCM $4 \times \text{CO}_2$ simulation at the four selected time periods and in the moist energy balance model simulations meant to reproduce the GCM $4 \times \text{CO}_2$ simulation at each of those time periods, as indicated by the text in panel (a). Panels from top to bottom show the temperature (a) raw (in units Kelvin), (b) mean-normalized (unitless), (c) mean-normalized symmetric component (unitless), and (d) mean-normalized antisymmetric component (unitless). Note differing vertical axis spans in each panel.

the most negative values in the deep tropics to a local maximum in the mid-latitudes steepens with time. As noted above, the meridional gradient of the surface temperature

response, even when mean-normalized, also steepens with time, although this occurs primarily in the extratropics.

The largest difference between successive time periods in the mean-normalized, symmetric component of the anomalous surface heat flux is between the years 1-10 and 21-100, when values in the mid-latitudes and even more so in the Tropics decrease considerably. Heat uptake by the oceanic subtropical cells, which links equatorial and subtropical SSTs on multi-decadal timescales (Burls and Fedorov 2014b; Fedorov et al. 2015), is a likely culprit — by years 21-100, the deep-tropical anomalous surface flux and its symmetric component are pointed into the atmosphere, and subtropical and equatorial surface warming is comparable (Fig. 5).

Unlike the radiative feedback parameter, whose anti-symmetric component starts fairly large and weakens with time, for the net surface flux the anti-symmetric component strengthens to some extent going from the first decade to the near-equilibrium response. Initially, the ocean is uptaking heat at essentially all latitudes, most of all in the Southern Ocean. By years 21-100, the anomalous surface flux in the deep tropics is pointed into the atmosphere, and this region spreads into the northern hemisphere subtropics in the subsequent centuries and millennia. At the multi-centennial and millennial timescales, a prominent tripole pattern emerges in the southern extratropics, with an anomalous net flux into the atmosphere in the mid-latitudes and over Antarctica and an anomalous net flux into the Southern Ocean in between. A similar, albeit weaker pattern emerges in the northern extratropics but with signs reversed, and these oppositely-signed tripole project strongly onto the anti-symmetric component. Nevertheless, as just described this behavior of the anti-symmetric fields does not appreciably influence the symmetric warming field.

5. Results from perturbed-cloud-albedo simulations

a. Simulations description

The perturbed-cloud-albedo simulations of Burls and Fedorov (2014b) include the aforementioned Pliocene-like simulation (denoted Experiment #16 of that study; see also Burls and Fedorov 2014a). Burls and Fedorov (2014b) describe them thoroughly. In brief, the simulations are run with preindustrial atmospheric composition (including CO_2) but with the liquid water path (LWP) — and for low-latitude perturbations the ice water path (IWP) also — artificially multiplied by a specified factor at each timestep, but only with respect to calculating shortwave radiative transfer (i.e. the longwave radiative transfer, dynamical, and hydrological calculations all use the unmodified value). In the subset of the Burls and Fedorov (2014b) simulations we analyze, these multiplicative factors are constant in time but take two distinct values, one within

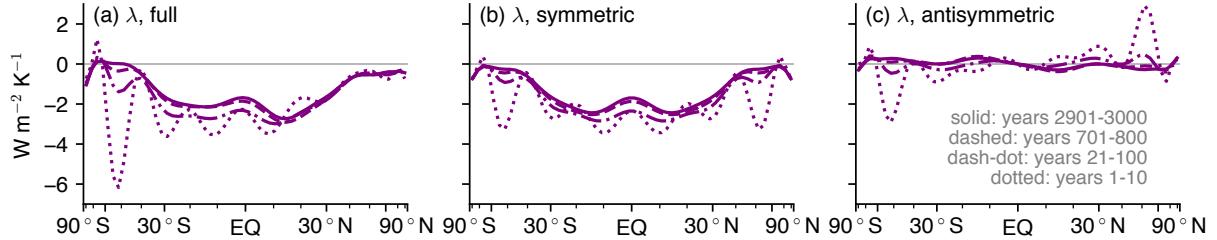


FIG. 7. Radiative feedback parameter used for the moist EBM simulations for each time period according to the text in panel c. From left to right, panels show the full field, its hemispherically symmetric component, and its hemispherically asymmetric component, all in units $\text{W m}^{-2} \text{K}^{-1}$. The full fields in panel (a) are used in the “full” MEBM simulations, and the symmetric fields in panel (b) are used for the symmetric-inputs MEBM simulations.

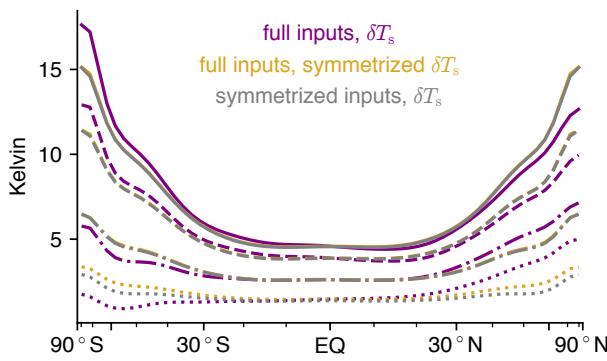


FIG. 8. Surface air temperature response in the MEBM simulations corresponding to each of the four time periods of the GCM $4 \times \text{CO}_2$ simulation (purple curves), along with the hemispherically symmetric component thereof (yellow curves) and the temperature responses in the MEBM simulations forced with hemispherically symmetrized fields rather than the full fields (grey curves). Time periods correspond to the line styles as specified in Fig. 6(a). Note that, except for the first decade (dotted curves), the yellow curves are hardly visible, because they are nearly identical to the grey curves plotted above them.

15°S - 15°N and another poleward thereof in both hemispheres. Moreover, their influence on a particular column at a particular timestep depends on the local cloud cover, with the calculations wholly unmodified under cloud-free conditions.

With respect to these simulations, we use the term “forcing” in a loose sense to refer to the magnitude and sign of the imposed modifications to cloud liquid and ice water paths in the shortwave radiation code. This is most directly related to the instantaneous forcing (e.g. Hansen et al. 1997). The effective radiative forcing (e.g. Tang et al. 2019) at any latitude can differ from this due to non-surface-temperature-mediated, rapid adjustments of the atmosphere to the forcing (e.g. Previdi et al. 2020). On the other hand, at all four timescales the TOA albedo anomaly and, except poleward of $\sim 70^{\circ}\text{S/N}$, net TOA radiative imbalance agree in sign with the imposed albedo modifica-

tions at all latitudes, with a step change at the 15°S/N transition (not shown; see Fig. 3a,b of Burls and Fedorov 2014b, for the TOA radiative imbalance for years 701-800). This suggests that rapid adjustment processes do not lead to qualitative differences between the instantaneous and effective radiative forcings.

Using the same numerical labeling convention as Burls and Fedorov (2014b), we analyze Experiments #2-17 (#1 is the preindustrial control), and they can be categorized into four distinct sets based on the relative signs and magnitudes of the shortwave cloud albedo perturbations imposed in the low- vs. high-latitude bands. Each set will be described in the subsequent subsections as they become relevant.

Because the CO_2 concentration remains preindustrial in all of the modified albedo cases, any global-mean temperature change results from changes in planetary albedo. Thus, the perturbed- CO_2 and perturbed-cloud-albedo and sets of simulations differ from one another in two primary ways: the former is subject to LW forcing that is relatively uniform, the latter subject to SW forcings with pronounced meridional structure. This is shown schematically in Fig. 10.

b. Extratropical cloud dimming and cloud brightening simulations

In Experiments #2-5, I/LWP is unchanged in the low-latitude band and LWP is increased in the extratropical bands by 80, 60, 40, and 20% respectively in #2 through #5, yielding global-mean cooling. Experiments #6-9 are analogous but with the extratropical LWP decreased rather than increased, by 20, 40, 60, and 80% respectively in #6 through #9, yielding global-mean warming. Comparing the equal-but-opposite perturbation pairs from #2-5 and #6-9 (i.e. #2 with #9, #3 with #8, #4 with #7, and #5 with #6) therefore reveals any nonlinear dependence of the response on the sign of the imposed I/LWP perturbation.

The top row of Fig. 11 shows the zonal-mean surface air temperature responses averaged over years 1-10, 21-100,

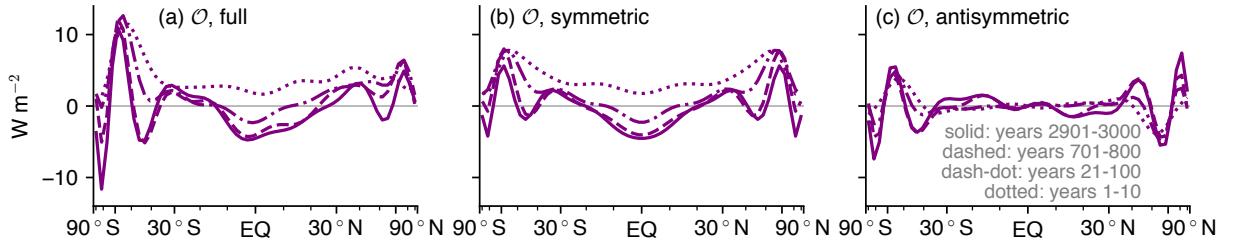


FIG. 9. Same as Fig. 7, but for the net surface flux, \mathcal{O} , in units W m^{-2} , signed positive downward.

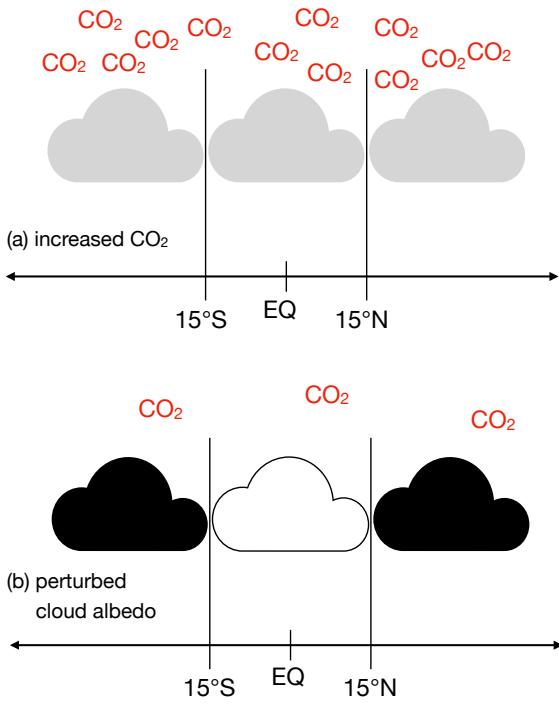


FIG. 10. Schematic depiction of the modifications imposed in the increased CO_2 simulations and the perturbed cloud albedo simulations. The number of red CO_2 symbols represents the CO_2 concentration, with three symbols corresponding to the control, and the darkness of the cloud symbols representing the imposed cloud albedo modification, with light gray corresponding to the control. In (a), CO_2 concentration has been increased, while the cloud albedo is unmodified. In (b), the CO_2 concentration is unmodified but the cloud albedo is decreased in the band 15°S - 90° in both hemispheres and increased equatorward thereof.

and 701-800 in the perturbed-cloud-albedo simulations #2-9. For the mean-cooling cases (#2-5), non-normalized responses are shown with sign reversed to facilitate comparison with the warming and CO_2 cases. Like the CO_2 cases, these perturbed cloud albedo cases exhibit polar-amplified temperature changes that are initially larger in the Arctic than Antarctic. Unlike the CO_2 cases or the

cloud dimming cases, the cloud brightening cases develop a prominent peak (in cooling) south of the North Pole beginning in years 21-100. An analogous feature exists in the Southern Hemisphere in essentially all simulations (including the perturbed- CO_2 simulations) and timescales beyond the first decade, but this northern manifestation is unique to the cooling simulations.

At each specified magnitude of the imposed LWP modification, cloud dimming yields more warming than brightening yields cooling — the peak magnitude of zonal-mean temperature change ranges from ~ 4 to 35 K in the cloud dimming simulations vs. ~ 3 to 12 K in the cloud brightening simulations. In both the first decade and years 21-100, all of the cooling cases are closest in magnitude to the weakest warming case (#6) except for the northern high-latitudes which come nearer to the next warming case (#7). In years 701-800, the relative Antarctic response strengthens for the cooling cases, with the strongest cooling case's peak near $\sim 60^\circ\text{N}$ slightly exceeding the second-strongest warming case locally. In the Antarctic, by contrast, the strongest cooling case only comes into the vicinity of the third-strongest warming case.

In both the brightening and dimming cases and at all timescales, the surface temperature response is single-signed across latitudes at all timescales, including the deep tropical latitudes (15°S - 15°N) in which clouds were not externally modified. This is not exactly surprising given the dynamical processes in the low latitudes that inhibit horizontal temperature variations in the free troposphere; as is commonly seen these free-tropospheric constraints impart their influence downward to the surface also. It nevertheless arises despite an instantaneous radiative forcing in these low latitudes that is exactly zero (and effective radiative forcing that is likely small as discussed above). This implicates anomalous energy transport convergence in the warming cases and divergence in the cooling cases (not shown).

The bottom row of Fig. 11 shows the global-mean-normalized surface air temperature responses in these simulations. The mean-normalized responses of the cloud-dimming simulations #6-9 resemble those of the increased- CO_2 simulations: Arctic warming is pro-

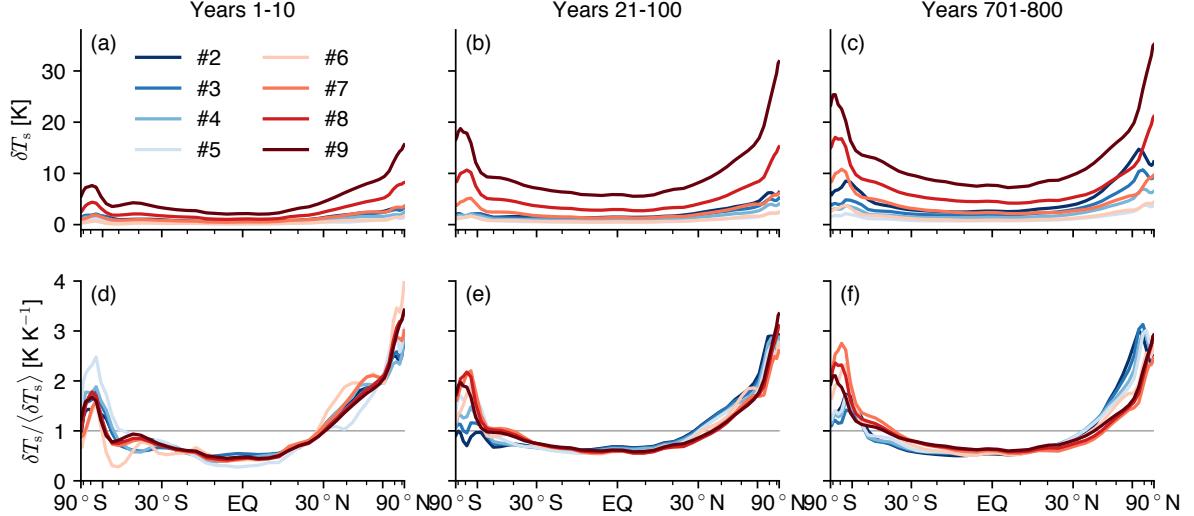


FIG. 11. Annual-mean, zonal-mean surface air temperature response for perturbed-cloud-albedo Experiments #2-9 averaged over (columns, left to right) years 1-10, 21-100, and 701-800, either (top row, units Kelvin) not normalized or (bottom row, unitless) normalized by the global-mean surface air temperature response in that simulation and time period. The non-normalized responses of #2-5, in blue shades in the top row, are shown with their sign reversed to ease comparison with simulations #6-9.

nounced and Antarctic warming muted initially, but the two high-latitude signals become more similar in subsequent decades and centuries.

Fig. 12 shows the hemispherically symmetric and antisymmetric components of the global-mean-normalized, zonal-mean surface air temperature responses at each time period for simulations #2-9. The cloud brightening simulations are more symmetric than the dimming simulations in the first decade and, unlike the dimming or CO₂ simulations, become *more* antisymmetric thereafter. Of the dimming cases, there are a few simulations in which eventually Antarctic warming exceeds Arctic warming over a $\sim 15^\circ$ band, as occurs also in the $2\times$ and $4\times$ CO₂ simulations. But unlike the CO₂ cases, this signal is not monotonic with the forcing magnitude. The greatest relative Antarctic warming occurs for the second-weakest magnitude forcing case (#7), though in principle this (and other behaviors) could change over subsequent millennia as full equilibrium is approached.

Despite these opposite-signed tendencies of the anti-symmetric component in the brightening vs. dimming simulations, the symmetric components of the global-mean-normalized surface air temperature response collapse toward nearly a single curve, with values near ~ 0.5 in the deep Tropics, increases toward unity in the mid-latitudes, and steeper slopes poleward thereof to values near ~ 2 . The biggest discrepancies from this occur primarily in the first decade and for the smallest forcing magnitudes (relatively smaller response in the deep tropics and larger response in the subtropics). The aforementioned local extremum near $\sim 60^\circ\text{N}$ unique to the cooling cases, along

with a similar signal in the southern high latitudes in both cooling and warming cases, also generates a corresponding peak in the symmetric component for the cooling cases that is absent from the warming cases.

Overlaid in Fig. 12(a) is the average across the sixteen mean-normalized, symmetric temperature responses from the perturbed-CO₂ simulations shown in Fig. 4 (four simulations and four time periods per simulation). The average symmetric temperature component for CO₂ forcing stays relatively flat over a wider range of low latitudes than that of all of the simulations #2-9 and at all time periods. This is not entirely surprising given the lack of forcing at low latitudes in the perturbed-albedo simulations. Poleward of $\sim 60^\circ$, however, its slope with latitude is comparable to the perturbed-albedo simulations #2-9. For the antisymmetric component, in contrast the CO₂ average [overlaid in Fig. 12(b)] is near the middle of the spread across the perturbed-albedo simulations #2-9.

c. Forcings with different meridional structures

Experiments #10-13 repeat #6-9 but with an opposite-signed and two-times-larger-magnitude perturbation applied in the low-latitude band. The increased low-latitude I/LWP acts to cool the deep tropics, and this reduces global-mean warming compared to the corresponding simulation from #6-9. Experiments #14-17 repeat #10-13 but with four-times rather than two-times magnitudes in the 15-90°S/N bands, further reducing the mean warming compared to #6-9. The simulations with mean warming, #6-17, can thus be categorized in two ways. The first is according to the relative values of the imposed perturbations

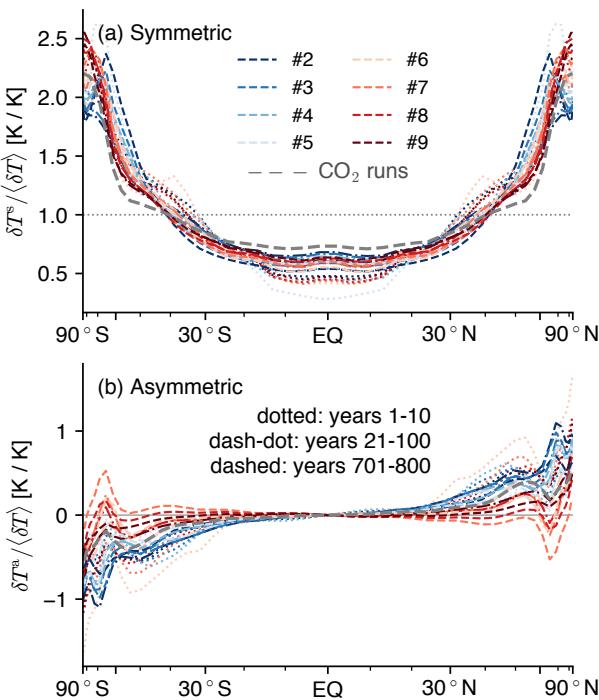


FIG. 12. As in Fig. 4, but for the perturbed-cloud-albedo experiments #2–9. Also included for comparison is the average across the four CO₂ values and four time periods. Unlike Fig. 11, the sign of #2–5 has not been reversed, since the cooling at each latitude normalized by a mean cooling yields a positive mean-normalized response at each latitude, matching the sign of the mean-warming cases #6–9.

in the tropical vs. extratropical bands, i.e. by the meridional structure of the imposed perturbation relative to the extratropical value. This yields the groupings #6–9 (0× in Tropics), #10–13 (−2× in Tropics), and #14–17 (−4× in Tropics). The second is in terms of the value of the imposed extratropical perturbation. This yields the groupings #6, 10, 14 (extratropical LWP -20%), #7, 11, 15 (-40%), #8, 12, 16 (-60%), and #9, 13, 17 (-80%). Within each of these triplets, the extratropical (instantaneous) forcing is fixed, but the opposite-signed tropical forcing, and with it the global-mean forcing and temperature response, differs. These two methods of comparison will prove useful in distinguishing between the roles of absolute vs. mean-normalized forcings.¹

¹These categorizations are not the same as the grouping of simulations by Burls and Fedorov (2014b) labeled Sets 1–4. Separately, Burls and Fedorov (2014b) perform additional simulations, labeled #18–21, in which the imposed increase of LWP/ILWP in the shortwave radiative transfer code spans 30°S–30°N and an 8-times larger magnitude imposed decrease spans 30–60° in either hemisphere, with no change applied poleward of 60°S/N. The zonal-mean surface air temperature response in these simulations is sufficiently complicated (not shown) that we do not analyze them in detail here.

In interpreting these results, it is useful to consider how these simulations would behave under two limiting cases, of negligible anomalous energy transport divergences on the one hand or of perfectly MSE-diffusive anomalous energy transports on the other. If perturbed energy transport divergences play no meaningful role, each column is effectively its own isolated system, and so the response at each latitude within the 15–90°S/N bands would be identical across the three simulations of each (#*n*, *n*+4, *n*+8) triplet, where *n* ∈ {6, 7, 8, 9}. Conversely, if the energy transport acts to diffuse the anomalous near-surface MSE field perfectly, then the response at each latitude effectively integrates the imposed forcing over all latitudes. We would then expect the surface temperature response patterns to depend solely on the global-mean forcing, which maps more onto the groupings #6–9, #10–13, and #14–17 with the same relative low- vs. higher-latitude perturbations than it does on the (#*n*, *n*+4, *n*+8) triplets with shared extratropical forcing values.²

The top row of Fig. 13 shows the raw zonal-mean surface air temperature change in simulations #6–17 for each of the three averaging periods. Line markings signify #6–9, #10–13, and #14–17 (solid, dashed, and dotted, respectively), and line colors signify which (*n*, *n*+4, *n*+8) triplet as indicated in the legend. In the extratropics, the raw temperature responses group together by high-latitude forcing more than by the meridional forcing structure in all averaging periods. Taking the Arctic in years 701–800 as an example, warming varies from ∼5 K in the (#6, 10, 14) triplet to ∼35 K in the (#9, 13, 17) triplet, with warming in individual simulations tightly clustered together within each triplet and well separated from any of the other simulations. This clustering is less tight in the Antarctic, but nevertheless the relative ordering of warming across the simulations is fixed at all latitudes poleward of ∼40°S/N. At low latitudes, by contrast, responses from different simulations do intersect each other, and within the highest-warming triplet (#9, 13, 17) equatorial warming ranges by ∼8 K.

All simulations #6–17 and time periods exhibit the familiar local maximum near ∼70°S and warming in the Northern Hemisphere that increases monotonically toward the north pole. In a minority of simulations in the first decade, fewer still in years 21–100, and #14 only in years 701–800, the increased 15°S–15°N cloud albedo is sufficiently strong to cool surface air temperature over roughly that band. Otherwise, despite the effective negative forcing locally the low latitudes warm, which must be due to anomalous energy transport convergence (again, provided

²This is imperfect, because the 15–90°S/N bands span a larger fraction of Earth's surface area (∼74%) than the 15–15°S/N band (∼26%). Thus, even if the actual forcing (which we do not explicitly compute) depended linearly on the imposed ILWP fractional modification, the global-mean forcing would not be identical within each of the three #6–9, #10–13, and #14–17 groups.

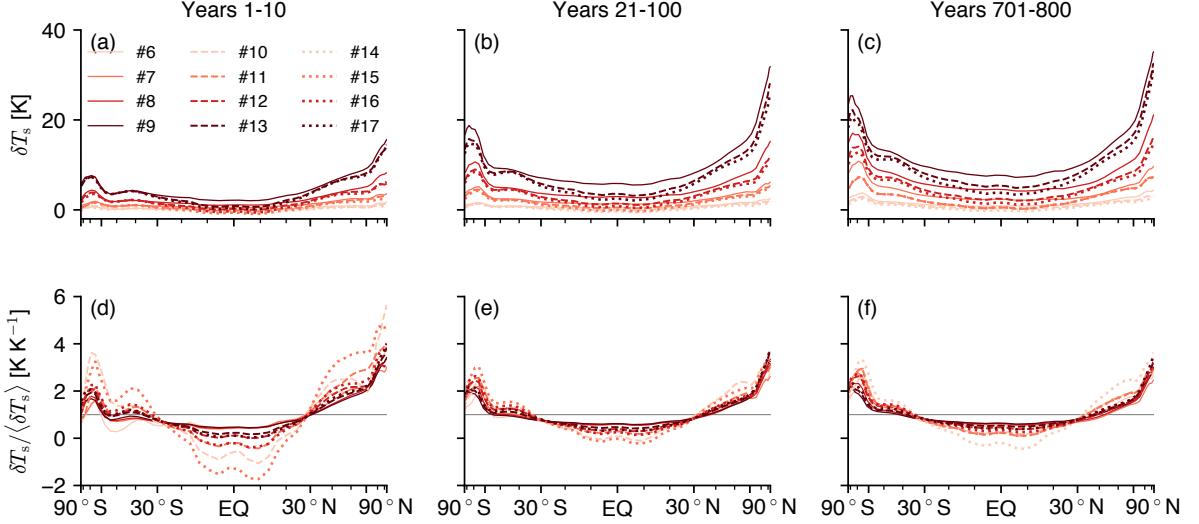


FIG. 13. As in Fig. 11, but for perturbed-cloud-albedo Experiments #6-17 according to the legend in panel (a).

rapid adjustments do not change the sign of the effective radiative forcing from its instantaneous value).

The bottom row of Fig. 13 shows the corresponding global-mean-normalized responses; #14 is excluded for years 1-10 and 21-100 for which its small global-mean response makes the normalization unmeaningful. Unlike for the raw responses, the 15-90°S/N forcing value that defines the $(n, n+4, n+8)$ triplets does not cleanly separate the responses. If anything, the clearest grouping to emerge is based on the spatial pattern of the forcing, with simulations #6-9 collapsing nearly identically onto one another in the Tropics and subtropics and to a lesser extent at higher latitudes. Particularly in the first decade, the same holds only somewhat for #10-13 and does not accurately describe #14-17. For the most part however the simulations do evolve in time toward a shared mean-normalized pattern, with outliers at each timescale primarily being those with the smallest mean warming.

Fig. 14 shows the hemispherically symmetric and antisymmetric components of the mean-normalized surface air temperature responses in Experiments #6-17; because of the large number of curves they are shown in separate panels for each averaging period rather than all together. In the first decade, the symmetric component displays a clear dependence on the mean, with the simulations with the weakest mean warming having the most polar-amplified normalized patterns. This weakens considerably but is nevertheless still evident over years 21-100 and 701-800. All of the responses are predominantly symmetric from the tropics through mid-latitudes, with the antisymmetric component being largest at each timescale for those with the smallest mean warming. And as in the other mean-warming runs already discussed, the degree of antisymmetry decreases with time. Although the symmet-

ric components of #6-17 do not collapse toward a single curve as much as the CO₂ simulations or the perturbed-albedo cases #2-9, curiously in all three time periods all of them nearly intersect around 30°S/N at a value of ~ 0.8 .

6. Conclusions

a. Summary

The zonal-mean surface air temperature response to abrupt CO₂ increases of 2, 4, 8, or 16× in 3,000-yr simulations performed in a low-resolution version of the CESM1 GCM exhibit an intriguing property: the hemispherically symmetric component, normalized by the global-mean, is quasi invariant across timescales and CO₂ levels. This includes the Arctic-amplified first decade after CO₂ is increased, when the whole northern extratropics, the Arctic especially, has warmed much more than the southern extratropics, the Southern Ocean especially. Over subsequent decades, centuries, and millennia, the damping effect of Southern Ocean downwelling on warming weakens relative to other processes, causing the antisymmetric warming component to steadily weaken in time. Conversely, after three millennia when the full climate system has reached (near) equilibrium — at which point, for context, zonal-mean warming exceeds 10 K at all latitudes and exceeds 30 K at the north pole — the mean-normalized, symmetric warming pattern has hardly varied from the value in its first decade common to all CO₂ values (with the first decade under 2×CO₂ a slight outlier).

A simple moist energy balance model prescribed with ocean heat uptake and radiative feedback fields inferred from four different time periods of the 4×CO₂ GCM simulation captures the GCM zonal-mean warming pattern

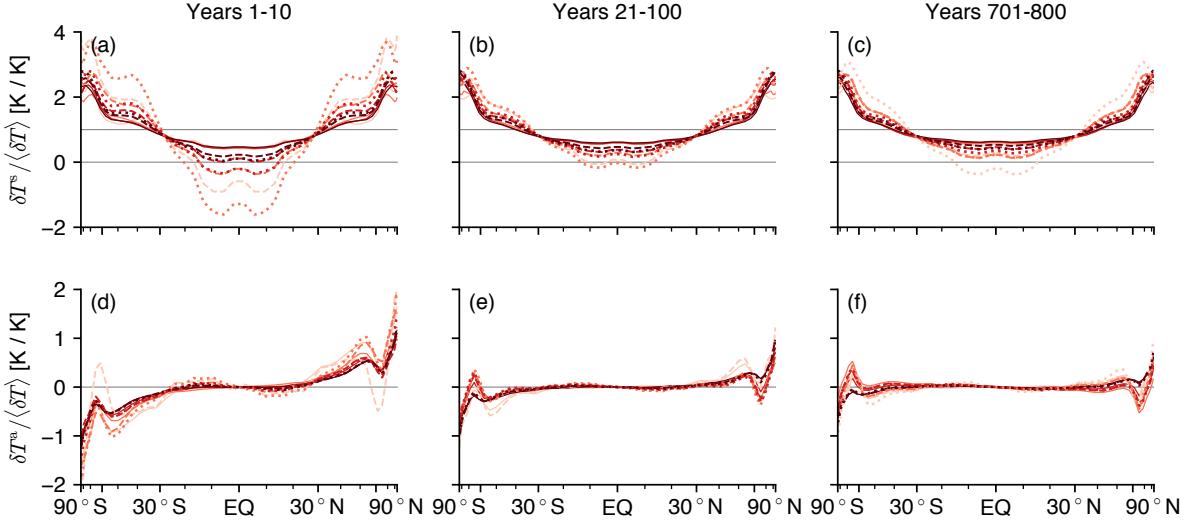


FIG. 14. As in Fig. 13, but for the hemispherically (top row) symmetric and (bottom row) antisymmetric components of the global-mean-normalized responses.

well overall, including the quasi invariance of the symmetric, mean-normalized component. This raises hope that the quasi invariance can ultimately be understood — though it remains incompletely so — through relatively straightforward process. Repeated simulations in the MEBM in which all inputs are first symmetrized yield warming patterns nearly identical to the symmetric component of the full simulations, suggesting a weak role for any rectifying interactions among antisymmetric components of any fields onto the symmetric component of surface warming.

In simulations in the same GCM with cloud albedo artificially modified in the extratropics (poleward of 15°S/N) and either unchanged or modified with the opposite sign in the deep tropics (15°S–15°N), for each individual simulation the hemispherically symmetric, mean-normalized warming response is again nearly invariant in time. Its meridional structure, however, depends on that of the imposed forcing, with the meridional slope in mean-normalized warming increasing with the tropical-extratropical applied albedo perturbation difference. In absolute terms (i.e. without normalization), high-latitude warming appears much more sensitive to the locally imposed forcing than to the meridional gradient thereof.

b. Discussion

In the perturbed cloud albedo simulations, the imposed forcings are strongly structured meridionally. But at least in terms of the percentage alterations to L/LWP in the SW radiative transfer calculations, they are hemispherically symmetric, and we expect that a properly computed measure of the corresponding adjusted radiative forcing would likewise be nearly hemispherically symmetric. In other

words, we have not examined any cases in which the imposed forcing itself exhibits a strong hemispherically antisymmetric component. It is quite reasonable to imagine that perturbations that are relatively more antisymmetric could yield mean-normalized symmetric warming patterns that differ substantially from what we're otherwise calling the quasi-universal pattern of our simulations.

Numerous questions remain. Of particular concern, does the quasi-invariance of the mean-normalized, symmetric component of surface warming across timescales and forcing magnitudes emerge in other GCMs under CO₂ forcing? For those models in which it does (including the MEBM), how strongly meridionally patterned or otherwise modified do imposed perturbations need to be for the mean-normalized, symmetric surface temperature response to vary markedly in time? Finally, what is a minimally complex, physically justifiable explanation for this quasi-invariance, preferably in the form of analytical expressions? We deem these questions worthy of further exploration.

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