

The Global Patterns of Instantaneous CO₂ Forcing at the Top of the Atmosphere and the Surface

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ABSTRACT: The radiative forcing of carbon dioxide (CO₂) at the top of the atmosphere (TOA) has a rich spatial structure and has implications for large-scale climate changes, such as poleward energy transport and tropical circulation change. Beyond the TOA, additional CO₂ increases downwelling longwave at the surface, and this change in flux is the surface CO₂ forcing. Here we thoroughly evaluate the spatiotemporal variation of the instantaneous, longwave CO₂ radiative forcing at both the TOA and surface. The instantaneous forcing is calculated with a radiative transfer model using ERA5 reanalysis fields. Multivariate regression models show that the broadband forcing at the TOA and surface are well predicted by local temperatures, humidity, and cloud radiative effects. The difference between the TOA and surface forcing, the atmospheric forcing, can be either positive or negative and is mostly controlled by the column water vapor, with little explicit dependence on the surface temperature. The role of local variables on the TOA forcing is also assessed by partitioning the change in radiative flux to the component emitted by the surface versus that emitted by the atmosphere. In cold, dry regions, the surface and atmospheric contribution partially cancel out, leading to locally weak or even negative TOA forcing. In contrast, in the warm, moist regions, the surface and atmospheric components strengthen each other, resulting in overall larger TOA forcing. The relative contributions of surface and atmosphere to the TOA forcing depend on the optical thickness in the current climate, which in turn is controlled by the column water vapor.

KEYWORDS: Anthropogenic effects/forcing; Climate change; Energy budget/balance; Greenhouse gases; Radiative forcing

1. Introduction

The radiative forcing of carbon dioxide (CO₂) at the top of the atmosphere (TOA) plays a central role in quantifying climate change and its global-mean value is a key aspect of radiative feedback analysis. Beyond the global mean, it has been recognized that the TOA radiative forcing of CO₂ is spatially inhomogeneous because of its relation to local atmospheric conditions (e.g., Feldl and Roe 2013; Zhang and Huang 2014; Huang et al. 2016, 2017; Jeevanjee et al. 2021). Previous literature has shown that the forcing distribution associated with current climate can drive changes in the mean tropical circulation (Merlis 2015; Shaw and Tan 2018) and poleward energy transport (Huang et al. 2017), owing to the differential forcing that arises from the consequences of the climatological spatial structure of temperature, water vapor, and the cloud distribution on the forcing. In addition to altering Earth's energy balance at the TOA, CO₂ increases affect the downwelling longwave radiation at the surface (Collins et al. 2006; Feldman et al. 2015).

The difference between the TOA and surface forcing of CO₂ affects the *atmospheric* energy balance and this, in turn, constrains the hydrological cycle response on fast time scales when the surface warming is small (Allen and Ingram 2002;

Previdi 2010; Pendergrass and Hartmann 2014; Samset et al. 2016; Allan et al. 2020) and is an important energetic consideration in solar radiation geoengineering schemes (Seeley et al. 2021). Besides, the CO₂ forcing has implications on distinct aspects of climate change, such as the strengthening of the West African monsoon since the 1980s (Dong and Sutton 2015) and the weakening of tropical circulation (Chemke and Polvani 2021). The understanding of CO₂ atmospheric forcing also helps distinguish disparate climate responses to aerosol forcing (O'Gorman et al. 2012). Moreover, the traditional TOA-centered perspective has the limitation in the context of energy transport because it affects the combined ocean and atmosphere transport, so the additional information from surface forcing can be used to isolate the atmospheric component of the change in energy transport (Huang et al. 2017). A thorough understanding of CO₂ radiative forcing at both the TOA and the surface is therefore critical to important aspects of both regional and global climate change.

There are two schools of thought as to what give rise to instantaneous forcing: saturation¹ of absorption bands (Pierrehumbert 2010; Jeevanjee et al. 2021; Romps et al. 2022) and displacement

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¹ In this work, “saturation” is used in the context of radiation. When the radiative effect is saturated, the gas already absorbs all the radiation it can at its present concentration (the absorptivity ≈ 1), so the absorbed radiative fluxes are insensitive to increasing gas concentration. This is distinguished from the saturation of gas concentration, where gaseous and condensed matter are in thermodynamic equilibrium, such as the saturated water vapor pressure constrained by the Clausius–Clapeyron equation.

of the effective emission layer (Huang and Bani Shahabadi 2014; Dufresne et al. 2020). These two perspectives suggest that distinct components of the atmosphere are critical to the forcing: the saturation of absorption bands focuses on the magnitude of emission temperature, while the emission layer argument highlights the lapse rate around the emission height. Aside from different views of the relevant temperatures, the emission temperature itself results from underlying spectral variations, so that the quantitative connection between an atmospheric profile and the instantaneous forcing is challenging.

Recent work has attempted to analytically quantify the instantaneous CO₂ forcing at the TOA, with simplifications to keep calculations manageable: the wavelength-varying absorption spectrum and complications of water vapor have been simplified in spectrally averaged ways. For example, Jeevanjee et al. (2021) considered how the broadening of the absorption bandwidth blocks surface emission, allowing for a relationship between this change in bandwidth and the surface temperature that determines the TOA forcing in a dry atmosphere. With moisture, however, the broadened CO₂ absorption blocks emission from water vapor instead, which is hard to estimate by simple approximations and results in a nonnegligible bias in the forcing estimate in midlatitudes. In contrast, Dufresne et al. (2020) examined a realistic atmospheric profile and concluded that the emission layer displacement (a shift to higher elevations with increased CO₂) dominates the CO₂ forcing, and the amount of emission layer change is insensitive to water vapor amount. Yet, a prerequisite for emission layer displacement argument to hold is that the absorptivity is saturated, and it is therefore not applicable in the atmospheric window (Huang and Bani Shahabadi 2014; Dufresne et al. 2020). The bias would be even larger in the dry regions, where a considerable amount of outgoing longwave flux stems from the surface via the atmospheric window (e.g., Costa and Shine 2012).

Compared with the TOA forcing, CO₂ forcing at the surface has been less thoroughly investigated. Conceptually, one might expect the surface forcing follows a similar line of thought as the TOA forcing with competing perspectives on absorption change and emission layer displacement. However, there are difficulties that arise because the effective emission layer for downward flux is intertwined with absorption change. This stands in contrast to the upward flux at the TOA, where these two terms can be roughly separated and the effective emission can be approximated as the layer where the optical depth is unity (although the exact number depends on the choice of vertical coordinate; Jeevanjee and Fueglistaler 2020). At the surface, the majority of downward flux comes from the very bottom of the atmosphere with a decaying contribution from air above. In response to changes in CO₂ concentration, the e -folding length of the decaying contribution changes and this is sensitive to the detailed near-surface atmospheric structure. Freese and Cronin (2021) considered the broadening of absorption band, which only depends on CO₂ concentration, along with surface temperature to estimate the Antarctic surface forcing. This implicitly assumes a fixed effective emission layer. But we will later show that this method cannot explain the local minimum of surface forcing in the deep tropics, where the surface temperature can be locally warmer.

Shakespeare and Roderick (2021), in contrast, lumped all variables into a parameterized effective emission height, but it has limited application in moist regions. These suggest that the surface forcing requires an intrinsically different conceptual thinking compared to the TOA forcing.

Collectively, the current physical understanding of CO₂ forcing suggests it is a challenge to formulate analytic forcing estimates that can capture the range of climate states observed on Earth. This leaves semiempirical expressions, such as the multivariate regression model in Huang et al. (2016) that use convenient atmospheric variables to predict CO₂ forcing, as a valuable and practical approach to characterize and quantify the forcing's structure.

In this study, we provide an in-depth examination of CO₂ forcing pattern at the TOA, surface, and atmosphere, including their annual-mean pattern, seasonal cycle, and the controlling factors. We also quantify the relative contribution of the surface emission versus atmospheric contributions to the TOA forcing to shed light on the roles of absorption band saturation and emission layer changes. The paper will proceed as follows. Section 2 describes the radiative transfer model and dataset we use. In section 3, we document the complete instantaneous CO₂ forcing pattern and its temporal variability. Then, the multivariate regression models are used to identify the controlling factors of CO₂ forcing (section 4), extending the approach of Huang et al. (2016) to the surface and atmospheric forcing for a more comprehensive understanding. In section 5, we scrutinize the cause of TOA forcing from the spectral dimension, which offers insights into the relationship between CO₂ forcing at the TOA, surface, and atmosphere. Last, conclusions and discussion are given in section 6.

2. Model and data

The radiative fluxes are calculated by a standalone Rapid Radiative Transfer Model, a version that has been widely used in general circulation models (RRTMG; Iacono et al. 2008). The longwave component of RRTMG, ranging from 10 to 3250 cm⁻¹, computes the fluxes in 16 bands. The atmospheric data we use are taken from the hourly ERA5 reanalysis dataset of the European Centre for Medium-Range Weather Forecasts, while the skin temperature and surface pressure over land are replaced by the hourly ERA5-land dataset. The ERA5-land dataset is a downscaled land product produced by an additional simulation forced by near-surface ERA5 fields, with thermodynamic orographic adjustment of temperature (Hersbach et al. 2020). The well-mixed greenhouse gas concentrations of CO₂, CH₄, and N₂O are prescribed to 380, 1.797, and 0.323 ppmv, respectively, as in Huang et al. (2016). The calculations are repeated with doubled CO₂ concentration (760 ppmv), with all other variables unchanged, and the instantaneous radiative forcing is defined as the difference between these two computed fluxes. We focus on the longwave component of forcing, as the CO₂ forcing associated with the shortwave is rather homogeneous (Huang et al. 2016). The radiation fluxes are calculated every 2.5° × 2.5° grid box for every hour over the 19-yr period 2000–18. The adoption of ERA5-land is for better consistency between RRTMG-calculated fields and ERA5 radiation fluxes

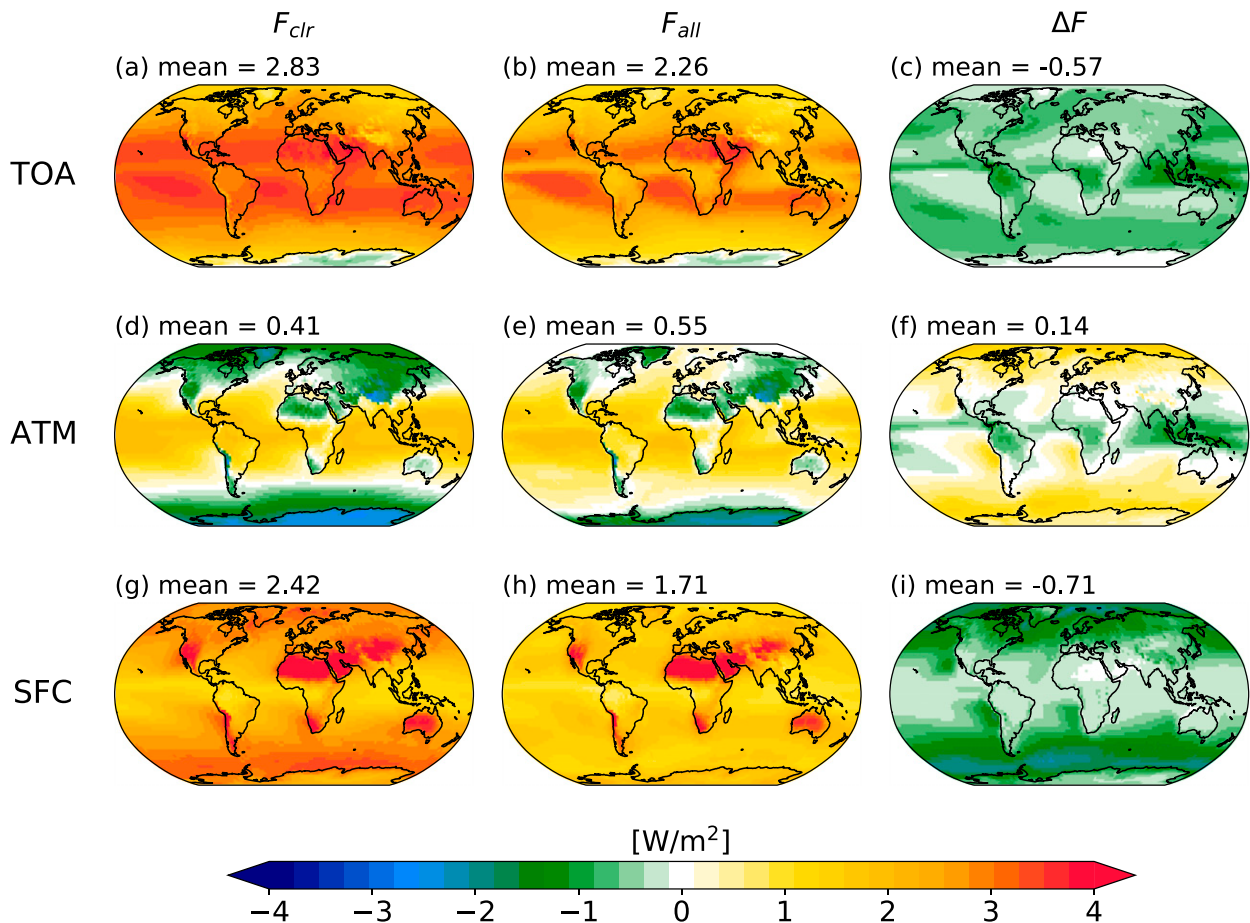


FIG. 1. The instantaneous radiative forcing of $2 \times \text{CO}_2$ (W m^{-2}). The forcing (a)–(c) at the TOA, (d)–(f) of the atmosphere, and (g)–(i) at the surface. (left) The clear-sky forcing (F_{clr}), (center) all-sky forcing (F_{all}), and (right) cloud masking effect ($\Delta F = F_{\text{all}} - F_{\text{clr}}$) on the forcing.

under current CO_2 concentration. As the difference in surface temperature is already reflected in the forcing calculation and regression models, the major conclusions remain the same.

We use ERA5 variables interpolated to 37 pressure levels as available from the CDS platform. The vertical structure of cloud cover, cloud liquid water content, and cloud ice water content from ERA5 are used for cloud inputs. At individual vertical levels, RRTMG calculates layer-wise cloud optical depth and radiative transfer in addition to clear-sky calculation. The all-sky fluxes at each level are the combined fluxes from the cloud and clear-sky calculation weighted by cloud fraction, which is treated with the maximum/random overlap scheme here. Our calculation on the cloud radiative effect is consistent with those available from ERA5, with correlation coefficients exceeding 0.99 at both the TOA and the surface (not shown). Although ERA5 is generally of superior quality of reanalyses in atmospheric state quantification (e.g., Wright et al. 2020), we acknowledge that the reanalysis is not the exact state of the atmosphere and there is uncertainty on the cloud radiative effect, so the calculated forcing could be affected. On the other hand, as we will show later, the forcing modified by clouds is well correlated to the cloud radiative

effect of the input atmosphere, so the regression models could also help evaluate the forcing uncertainty due to the effect of climatological clouds.

3. The forcing pattern and variation

a. Spatial variation: Climatology

The 19-yr average of instantaneous CO_2 forcing pattern is shown in Fig. 1. Under clear sky, the global-mean forcing is 2.83 W m^{-2} at the TOA and 2.42 W m^{-2} at the surface, resulting in a convergence of 0.41 W m^{-2} within the atmosphere: this difference defines the atmospheric forcing. The forcing evolves with the base climate during the course of 19 years. From 2000 to 2018, the annual-mean TOA forcing varies between 2.80 and 2.86 W m^{-2} and the surface forcing varies between 2.44 and 2.40 W m^{-2} , with no clear trend during the period. The TOA and surface forcing do not always covary, so the magnitude of the atmospheric forcing fluctuates between 0.36 W m^{-2} in 2000 and 0.46 W m^{-2} in 2018. The clouds decrease the global-mean TOA and surface forcing by 0.57 and 0.71 W m^{-2} , respectively, and they result in an increase of the all-sky atmospheric forcing by 0.14 W m^{-2} .

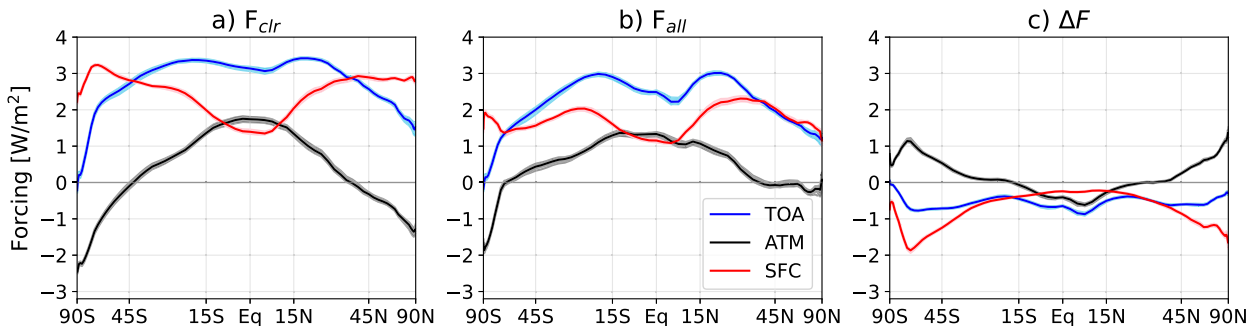


FIG. 2. The zonal-mean of annual-mean instantaneous forcing for (a) clear-sky (F_{clr}) and (b) all-sky (F_{all}) conditions, and (c) the cloud masking effect ($\Delta F = F_{\text{all}} - F_{\text{clr}}$). The thick line is the 19-yr climatology, and the annual means of individual years are plotted in a lighter color.

Compared with Pincus et al. (2020), who sampled a subset of re-analysis profiles with line-by-line radiative transfer calculations, the magnitude of clear-sky, global-mean TOA forcing is similar to documented 2.71 W m^{-2} in their study, yet the surface forcing we report here is 20% larger than their results. Another similar calculation is that of Huang et al. (2017), who calculated an all-sky forcing based on a global climate model's atmospheric state of 2.6 W m^{-2} at the TOA and 1.4 W m^{-2} at the surface, meaning a difference of around 15% at the TOA and 30% at the surface. Although these numbers are not directly comparable as a result of several factors, such as difference on the choice of base climate, initial CO_2 concentration, sampling location, and radiative transfer model used, these results together provide an indication of the variability in the range of forcing values.

Figure 1 shows that the forcing pattern displays a first-order dependence on latitude, especially for the clear sky (F_{clr} ; left column). The clear-sky TOA forcing decreases from the low to high latitudes, ranging from 3.4 to 1.5 W m^{-2} with a weak but discernible local minimum in the deep tropics, where the forcing slightly decreases to 3 W m^{-2} . This relatively small forcing in the moist region has been identified as the water vapor masking effect (Huang et al. 2016; Jeevanjee et al. 2021) because the absorption by extra CO_2 , which centers around $15 \mu\text{m}$, is reduced by the rotation band of pre-existing water vapor. The forcing in polar regions is strongly hemispherically asymmetric and is negative in the Antarctic, consistent with previous work (Zhang and Huang 2014; Schmithüsen et al. 2015; Huang et al. 2016). The clear-sky surface forcing exhibits a local minimum in the tropics like the TOA component but is more prominent, owing to the stronger water vapor masking effect. Additionally, the surface forcing is more zonally asymmetric, with amplified forcing over dry regions and mountains (the Sahara Desert, Australian deserts, the Rockies, and the Andes) that exceeds 4 W m^{-2} in the annual mean. The resulting atmospheric forcing enhances the equator-to-pole energy gradient due to it being positive at the equator and decreasing poleward. We also note that in dry regions, the strong surface forcing exceeds the TOA forcing and creates zonally anomalous negative atmospheric forcing.

The all-sky forcing (F_{all}) is shown in Fig. 1 (middle column). The all-sky TOA forcing shows a clear trace of the tropical rain belt with locally small forcing, suggesting a stronger masking

effect by water vapor and clouds combined. The clouds also reduce forcing in the extratropics by 0.6 W m^{-2} . Compared with the TOA forcing, the surface forcing shows even larger regional forcing deviation from the global mean in the deserts and major mountain ranges where clouds rarely present. The latitudinal forcing dependence is largely eliminated by low clouds in the extratropics. In all-sky atmospheric forcing, one can also see that the clouds impose a nonnegligible masking effect that smooths the meridional forcing gradient, with a reduced forcing peak around the equator and an overall increased forcing in the extratropics due to the surface forcing decrease. The strong negative clear-sky atmospheric forcing in the Arctic is also largely counteracted by clouds, bringing the annual-mean forcing close to neutral there.

Although the clouds themselves are assumed unchanged in this study, they impose large cloud masking effects (ΔF), defined as the forcing difference between the all sky and clear sky (Fig. 1, right column). Clouds reduce the CO_2 forcing at the TOA and surface. The clouds affect radiation with multiple factors, including different cloud height, fraction, and thickness, which modify the optical structure of the atmosphere (Zelinka et al. 2012). As the clouds are mostly optically thick, the radiative fluxes below (above) the clouds are hard to transmit to the TOA (surface). Therefore, the cloud type that influences the TOA forcing the most would be high clouds, since there is a large portion of the atmospheric column beneath clouds that no longer contributes to the forcing. The same principle is applicable to the surface forcing, for which it is the low clouds that matter. These arguments are supported by Fig. 1. At the TOA (Fig. 1c), the cloud effect is particularly strong in the deep tropics, where the thick and high cumulus from deep convection is more frequent. In contrast, the cloud masking effect for the surface forcing (Fig. 1i) is found amplified in the Arctic and the Southern Ocean, where widespread low clouds exist. In brief, the presence of clouds effectively limits the atmosphere column that is subject to CO_2 change and therefore reduces the forcing magnitude. In section 4, we will show that the cloud masking effect can be well estimated by the climatological cloud radiative effect (Zhang and Huang 2014).

Figure 2 shows the zonal means of the annual-mean radiative forcing maps of Fig. 1, with interannual variations shown

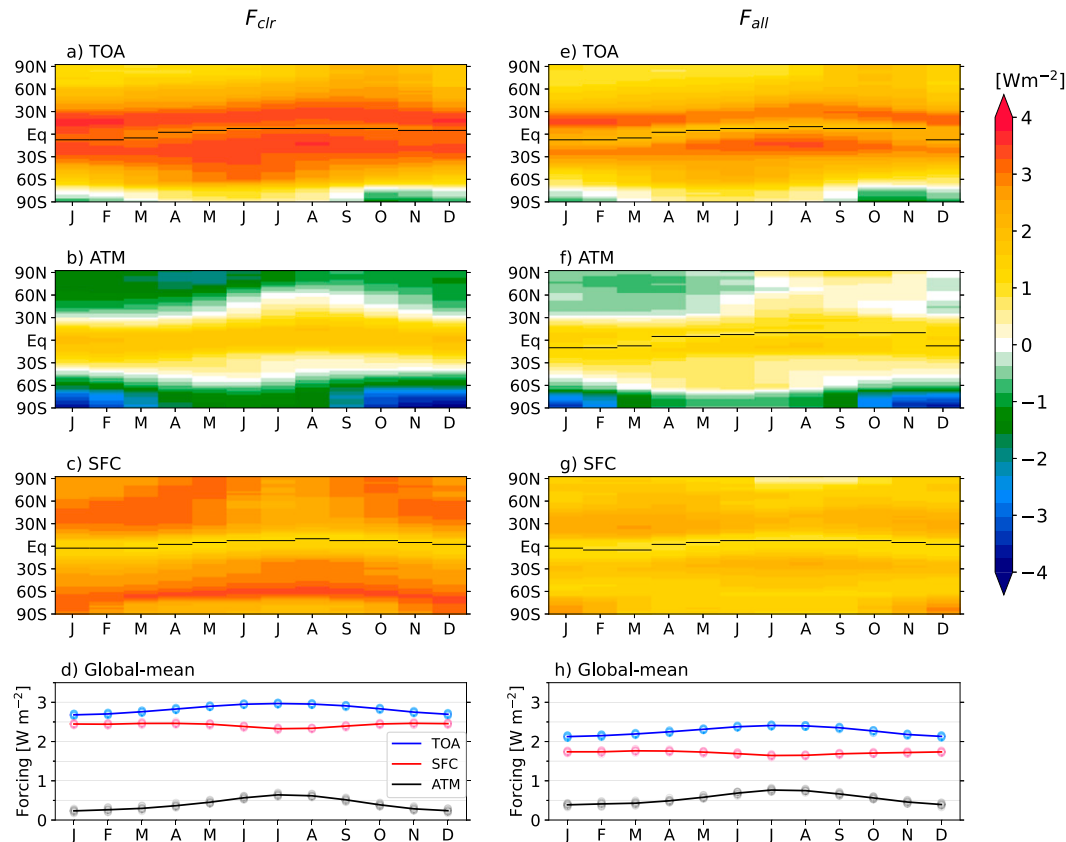


FIG. 3. The seasonality of climatological monthly average of zonal-mean clear-sky radiative forcing at the (a) TOA, (b) atmosphere, and (c) surface and the all-sky radiative forcing at the (e) TOA, (f) atmosphere, and (g) surface. The black horizontal bar in (a), (c), and (e)–(g) denotes the latitude with forcing minimum in the low latitudes (15°N–15°S). Also shown is the seasonality of global-mean forcing for (d) clear sky and (h) all sky, with each circle showing the global mean of monthly mean forcing in different years for the forcing at the TOA (blue), atmosphere (black), and surface (red), with the 19-yr averaged forcing connected in lines. Note that (a), (c), (e), and (g) are plotted with another color bar for better visualization in Fig. S1 in the online supplemental material.

in a lighter color. For both clear sky and all sky, the surface forcing is relatively symmetric between the hemispheres. The TOA component has strong hemispheric asymmetry in the polar regions, with the forcing in the Arctic larger than in Antarctica by 1 W m^{-2} . For clear sky, the atmospheric forcing peaks exactly at the equator, a consequence of cancelling local minima of both the TOA and the surface forcing that are located northward of the equator. For all sky, the atmospheric forcing in the Northern Hemisphere is greatly enhanced in the polar region and does not monotonically decrease poleward; it becomes rather close to zero between 35° and 90°N, primarily due to reduced surface forcing in the higher latitudes. Additionally, the latitude that the atmospheric forcing changes sign in the Southern Hemisphere shifts poleward from 45°S for clear sky to 60°S for all sky. Clouds also exert notable changes in the low latitudes, especially for the tropical TOA forcing. Within the tropics, the clouds lower the TOA forcing more than the SFC forcing and further mitigate the meridional forcing variation, with a greater forcing decrease around the annual-mean intertropical convergence zone (ITCZ)

for the TOA and consequently smaller atmospheric forcing. The atmospheric forcing no longer peaks at the equator, and it instead has a broad local maximum between 0° and 15°S. The interannual variability of the zonal- and annual-mean forcing is weak (shading in Fig. 2). The largest variability is in the Arctic, where the all-sky atmospheric forcing oscillates about zero with a range of 0.5 W m^{-2} . While the clouds damp the meridional atmospheric forcing gradient, we note that this should be distinguished from the cloud effects on radiative feedback, which may result in different meridional gradients (e.g., Zelinka and Hartmann 2010; Huang et al. 2017).

b. Seasonal variation

The instantaneous forcing also exhibits remarkable seasonality, as displayed in Fig. 3. The seasonality of CO_2 forcing is characterized to provide a comparison with the more pronounced seasonality of anthropogenic aerosol (e.g., Zhang et al. 2005) and solar forcing geoengineering schemes (e.g., Govindasamy and Caldeira 2000), which have well-established seasonality that results from the temporal variations in aerosol

concentration and insolation. The clear- and all-sky forcing, particularly in low latitudes, show similar seasonal dependence in both hemispheres. For instance, the TOA and surface forcing in the subtropics are amplified in the wintertime and the local minimum migrates with the ITCZ, resulting in a weak movement of atmospheric forcing peak. The latitudinal extent of positive atmospheric forcing, however, does not simply synchronize with the tropical rain belt movement, especially when the cloud effect is considered. For example, the northernmost latitude that the atmospheric forcing changes sign falls in August, late boreal summer, for clear sky, and under all sky the positive forcing blob can extend to the Arctic in boreal fall (August–November). In comparison, the southernmost latitude with positive atmospheric forcing occurs in June, the austral winter, for both clear sky and all sky.

The polar region shows even more hemispheric asymmetry for the TOA and the atmospheric components when analyzed seasonally. In particular, the TOA forcing in the Antarctic is negative from October to February, while it is always positive in the Arctic. The surface forcing is consistently positive and varies in an opposing manner to the TOA counterpart, which enhances the atmospheric forcing's seasonal variation. The seasonality of global-mean TOA and surface forcing (Figs. 3d,h) is opposite as well: the TOA forcing maximum falls in July, which is exactly the time with surface forcing minimum. The all-sky forcing has seasonality similar to the clear-sky forcing, implying that change in forcing from cloud masking ΔF is rather constant temporally in the global average. In the next section, we will show that the seasonality of forcing is related to the climatology of humidity and surface temperature, whose global means are greatest in July.

4. Regression model

In the previous section, we showed that the TOA, surface, and atmospheric forcing from a uniform doubling of the CO₂ concentration are highly inhomogeneous spatially and temporally. To further understand the dependence of the forcing pattern on the atmospheric and surface state, we begin from a statistical perspective following the multivariate regression analyses in Huang et al. (2016). In particular, we extend the regression model to the surface forcing. An expression for the atmospheric forcing can then be formed by the linear combination of the TOA and surface regression models. In this way, the CO₂ radiative forcing can be estimated directly from the local distributions of geophysical variables. This method also quantifies the sensitivity of the radiative forcing to the predictor variables, revealing their relative importance.

The form of the multivariate regression model is

$$\hat{F} = F_0 + \sum_{i=0}^n A_i \frac{y_i - y_{i0}}{y_{i0}}, \quad (1)$$

where $\hat{F}$ is the forcing prediction, and the subscript 0 denotes the multiyear global average. The regression coefficients A_i are determined by the area-weighted least squares regression method. The variables y_i are chosen geophysical variables

that serve as predictors normalized by its global mean y_{i0} . As the predictors y_i are normalized, the dimensions of $\hat{F}$, F_0 , and A_i in Eq. (1) are all in units of W m^{-2} . The regression coefficients are obtained from linear fit of global, monthly mean data in the 19-yr span and are constant in time (i.e., there is no seasonal dependence of the coefficients). This regression model, therefore, accounts for the spatial and temporal variations about the global mean, which is specified.

The central question is then how to choose predictors that are physically meaningful and easily accessible. Huang et al. (2016) identified several key variables that well predict the clear-sky TOA forcing, including the surface temperature (T_s), stratospheric temperature at 10 hPa (T_{10}), and column water vapor (CWV; in kg m^{-2}). The all-sky forcing is obtained by adding the cloud effect, which is predicted by the cloud radiative effect (CRE; in W m^{-2}) of the current climate (Zhang and Huang 2014; Huang et al. 2016). The physical motivation for each of these is as follows: T_s captures the equator-to-pole thermal emission pattern that explains a large portion of forcing gradient; T_{10} identifies the surface–stratosphere temperature contrast, which is pertinent to the emission layer elevation within the stratosphere, the major mechanism of longwave flux change in the CO₂ absorption band [Huang et al. 2016; see also Fig. 1a of Jeevanjee et al. (2021)]. The water vapor usually damps the forcing by reducing the energy that additional CO₂ can absorb because the rotation band of H₂O molecules partially overlaps with CO₂ absorption in the far infrared (referred to as water vapor masking of the CO₂ forcing). A similar argument holds for the cloud effect on the CO₂ forcing, which exhibits a wider absorption spectrum that is highly correlated to the current cloud radiative effect.

Narrowing the choice of predictors to those that are commonly available in the reanalysis and climate model output, we find that the square of column water vapor (CWV²) is essential to predict the surface forcing well, in addition to those variables that are already identified for the TOA forcing. The adoption of CWV² accounts for the pronounced water vapor absorption, such as the continuum absorption that is proportional to the square of concentration (Pierrehumbert 2010). The resulting regression models for clear-sky forcing $\hat{F}_{\text{clr}}$ are

$$\begin{aligned} \hat{F}_{\text{clr,TOA}} = & 2.83 + 15.85 \frac{T_s - 288.4}{288.4} - 8.85 \frac{T_{10} - 228.6}{228.6} \\ & + 0.47 \frac{\text{CWV} - 24.4}{24.4} - 0.43 \frac{\text{CWV}^2 - 846.2}{846.2} \quad \text{and} \end{aligned} \quad (2)$$

$$\begin{aligned} \hat{F}_{\text{clr,SFC}} = & 2.42 + 16.24 \frac{T_s - 288.4}{288.4} + 0.55 \frac{T_{10} - 228.6}{228.6} \\ & - 4.34 \frac{\text{CWV} - 24.4}{24.4} + 1.34 \frac{\text{CWV}^2 - 846.2}{846.2} \end{aligned} \quad (3)$$

for the TOA and surface forcing, with 98% and 91% variance explained, respectively. The atmospheric forcing is obtained by differencing the TOA and surface forcing:

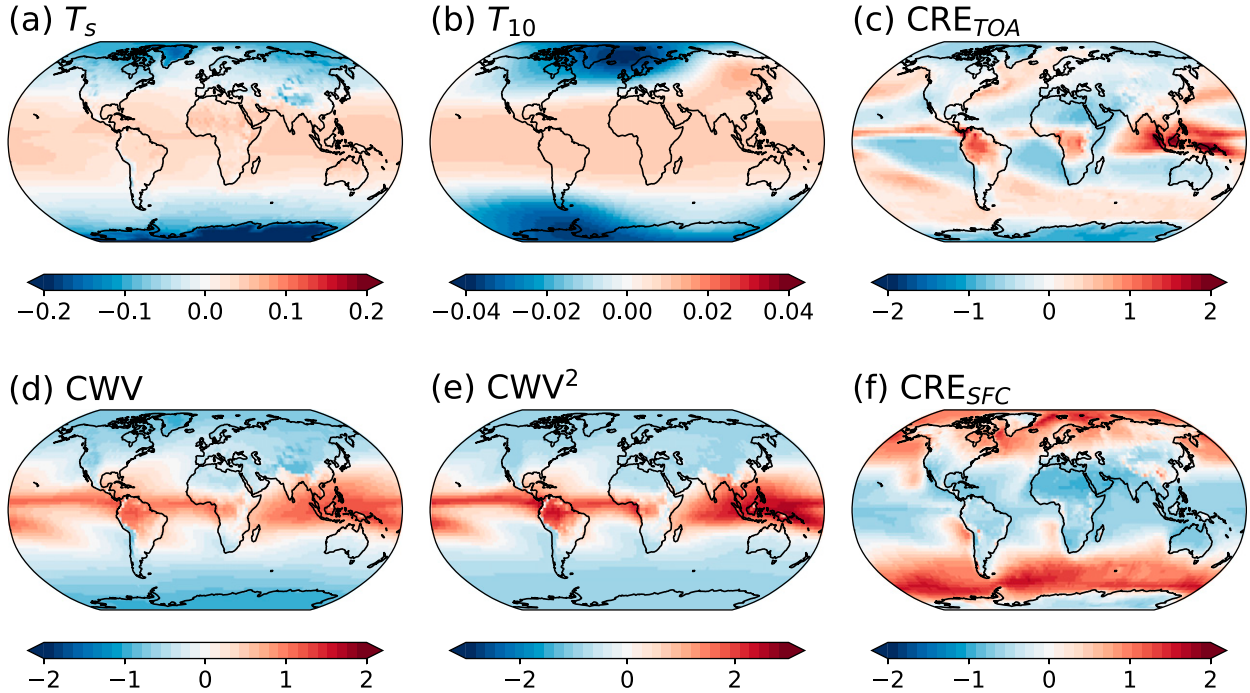


FIG. 4. The climatology of normalized predictors for multivariate regression models in Eqs. (2)–(10): (a) surface temperature, T_s ; (b) stratospheric temperature at 10 hPa, T_{10} ; (c) cloud radiative effect at the TOA, CRE_{TOA} ; (d) column water vapor, CWV ; (e) column water vapor squared, CWV^2 ; and (f) cloud radiative effect at the surface, CRE_{SFC} . The sign of column water vapor and column water vapor squared may differ because of the normalization.

$$\begin{aligned}\hat{F}_{\text{clr,ATM}} &= \hat{F}_{\text{clr,TOA}} - \hat{F}_{\text{clr,SFC}} \\ &= 0.41 - 0.39 \frac{T_s - 288.4}{288.4} - 9.40 \frac{T_{10} - 228.6}{228.6} \\ &\quad + 4.81 \frac{CWV - 24.4}{24.4} - 1.77 \frac{CWV^2 - 846.2}{846.2}.\end{aligned}\quad (4)$$

Although the regression models here include four variables for each, the TOA and surface forcing are well predicted by three predictors: T_s , T_{10} , and CWV for the TOA forcing, and T_s , CWV , and CWV^2 for the surface forcing. The fourth predictor is introduced to allow the atmospheric forcing to result from the difference of the TOA and SFC regression models. As the downward flux at the surface mainly stems from the emission in the lower troposphere with high water vapor concentrations, it is straightforward to expect a larger portion of forcing to be influenced by CWV at the surface than at the TOA. Similarly, the magnitude of the T_{10} coefficient in the surface forcing is less than 10% than that of the TOA forcing, suggesting a limited influence of stratospheric temperature at the surface.

The prediction for all-sky forcing is obtained by adding cloud masking effects on top of clear-sky equations:

$$\hat{F}_{\text{all,TOA}} = \hat{F}_{\text{clr,TOA}} + \widehat{\Delta F}_{\text{TOA}}, \quad (5)$$

$$\hat{F}_{\text{all,SFC}} = \hat{F}_{\text{clr,SFC}} + \widehat{\Delta F}_{\text{SFC}}, \quad (6)$$

$$\hat{F}_{\text{all,ATM}} = \hat{F}_{\text{clr,ATM}} + \widehat{\Delta F}_{\text{ATM}}, \quad (7)$$

where $\widehat{\Delta F}$ denotes the cloud masking effect, which is well correlated to the cloud radiative effect (Zhang and Huang 2014). Using the cloud radiative effect at the TOA and the surface as predictors, we find

$$\widehat{\Delta F}_{\text{TOA}} = -0.57 - 0.53 \frac{CRE_{\text{TOA}} - 20.06}{20.06}, \quad (8)$$

$$\widehat{\Delta F}_{\text{SFC}} = -0.71 - 0.78 \frac{CRE_{\text{SFC}} - 21.84}{21.84}, \quad (9)$$

$$\begin{aligned}\widehat{\Delta F}_{\text{ATM}} &= \widehat{\Delta F}_{\text{TOA}} - \widehat{\Delta F}_{\text{SFC}} \\ &= 0.14 - 0.53 \frac{CRE_{\text{TOA}} - 20.06}{20.06} \\ &\quad + 0.78 \frac{CRE_{\text{SFC}} - 21.84}{21.84}.\end{aligned}\quad (10)$$

Although simple in form, Eqs. (8) and (9) explain more than 97% of the variance for the cloud masking effects.

As the regression models explain more than 90% of forcing variance, the map of predictors reveals each variable's role in the forcing variability because the predicted forcing pattern can be viewed as the superposition of predictor contributions. The climatology of normalized predictors is shown in Fig. 4. The temperature predictors, T_s and T_{10} , vary mostly with latitude and have only slight zonal variation. This shows their

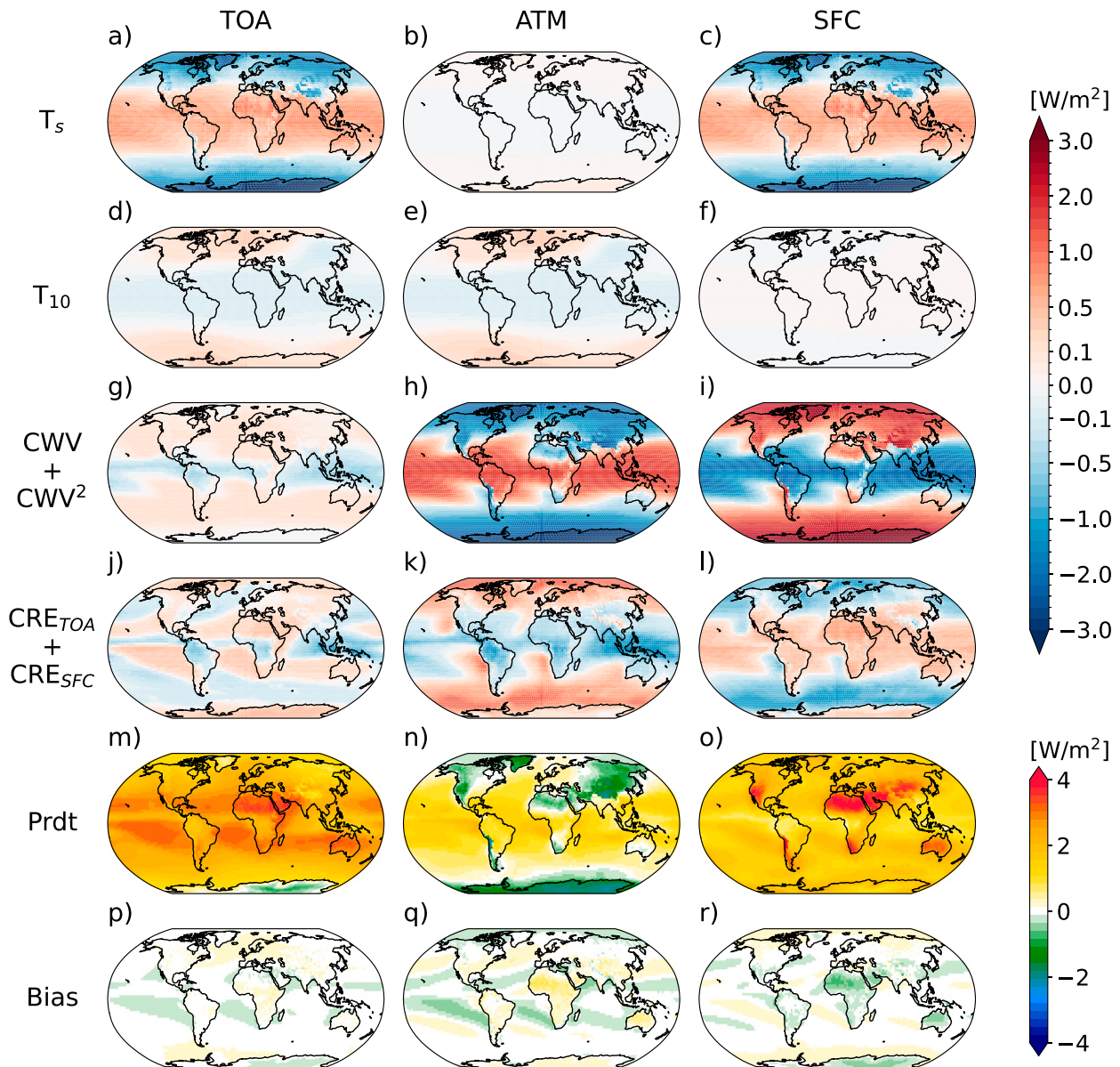


FIG. 5. (a)–(l) The partial contribution of multiyear mean forcing anomalies at the (left) TOA, (center) atmosphere, and (right) surface by multiplying the regression coefficients in Eqs. (2)–(10) with predictor anomalies. Note that the color intervals are not uniform. (m)–(r) The predicted all-sky forcing (the sum of top four rows plus the multiyear global-mean forcing) and the bias are shown.

role in explaining the meridional forcing gradient, consistent with the first-order dependence of the greenhouse effect on thermal emission. The strong zonal variation of CWV and CWV^2 in low latitudes accounts for the contrast between the dry, subsiding regions and moist, convecting regions within the tropics, a prominent aspect of the clear-sky surface forcing (Fig. 1). The cloud radiative effects at the TOA and surface reflect the high and low cloud cover, which are of large magnitude in the tropical ascending region and midlatitudes respectively (Figs. 4c,f). This structure resembles the cloud effect on forcing (Figs. 1c,f,i), consistent with the CRE explaining more than 97% of the forcing variance.

The contribution of individual components in the regression models is further visualized in Fig. 5 by combining the predictor patterns with coefficients of the regression models. For the TOA forcing, T_s is the dominant component, whereas T_{10} , CWV , and CRE all slightly offset it zonally and meridionally. The forcing dependence on temperature contrast between T_s and T_{10} can be viewed as a consequence of the “lapse rate” (Huang et al. 2016) or a “swap of surface and stratospheric emission” (Jeevanjee et al. 2021). The narrow belt with a local forcing minimum in the tropics is seen in CWV and CRE components, echoing the masking effect of water vapor and clouds (Merlis 2015; Huang et al. 2016). The

T_{10} also reduces the forcing in the tropics, but its effect is smaller than CWV and CRE. Interestingly, the negative forcing over the Antarctic seems well captured by extremely cold T_s (Figs. 5a,m), without knowledge of near-surface temperature inversion that has been identified as a factor accounting for the negative greenhouse effect (Flanner et al. 2018). We find that T_s alone reproduces much of the forcing variation temporally (Figs. 6a,c) because T_s dominates the surface–stratosphere temperature difference in most places. In Antarctica, however, T_{10} exhibits larger seasonal variation than T_s . The warmer T_{10} in Southern Hemisphere summer therefore contributes to negative TOA forcing from October to February [Eq. (2) and Fig. 3a].

For the surface forcing, the contribution by T_s is similar to the TOA prediction due to the similar magnitude of the regression coefficient, but the T_s component is offset by the substantial masking effect of CWV. CWV is a column-integrated quantity that is sensitive to the lower troposphere because of the bottom-heavy nature of specific humidity distribution, so one might intuitively expect a stronger CWV masking effect for surface forcing than the TOA. Of course, CWV may be closely related to T_s because the saturation specific humidity is governed by the Clausius–Clapeyron relation. However, the relative humidity can play an important role in shaping the zonal asymmetry of the forcing (Fig. 5i). Interestingly, unlike the TOA forcing where CWV and clouds show consistent signs of forcing anomaly in low latitudes, the pattern predicted by clouds is opposite to CWV because it is the low clouds rather than high clouds that are more relevant to the surface forcing (Figs. 4c,f). The cloud masking effect therefore more strongly occurs in the extratropics rather than the tropics for the surface forcing.

Although the atmospheric forcing is simply the forcing difference between two boundaries, some interesting results can be noted. First, the T_s contribution to the TOA and surface forcing almost cancel out, leaving this component even smaller than T_{10} [Eq. (4) and Figs. 5b,e]. Second, with strong CWV masking on the surface forcing, CWV is the dominant variable for the atmospheric forcing and accounts for the equator-to-pole forcing gradient (Fig. 5h). Moreover, as different cloud types exert distinct masking effects at the TOA and surface, the cloud masking effect does not simply cancel between the surface and TOA. Unlike the CWV masking, which actually produces the meridional forcing gradient within the atmosphere, the clouds overall mitigate the atmospheric forcing contrast both zonally and meridionally (Fig. 5k).

The seasonality of predicted global-mean forcing is displayed in Fig. 6. Overall, the picture of seasonality is in line with the results for the spatial pattern: the TOA forcing variation is best captured by T_s because T_s dominates the surface–stratosphere temperature difference, the surface forcing is dominated by CWV seasonality while being opposed by T_s , and the atmospheric forcing is well represented by CWV. One new dimension is that, for temporal variability, T_{10} contributes more to the TOA forcing than CWV and clouds. Also, the net cloud masking effect is quite steady seasonally. The largest bias in boreal summer can be traced back to larger underpredicted surface forcing in dry regions, including the Sahara Desert and Antarctica (Fig. 5q).

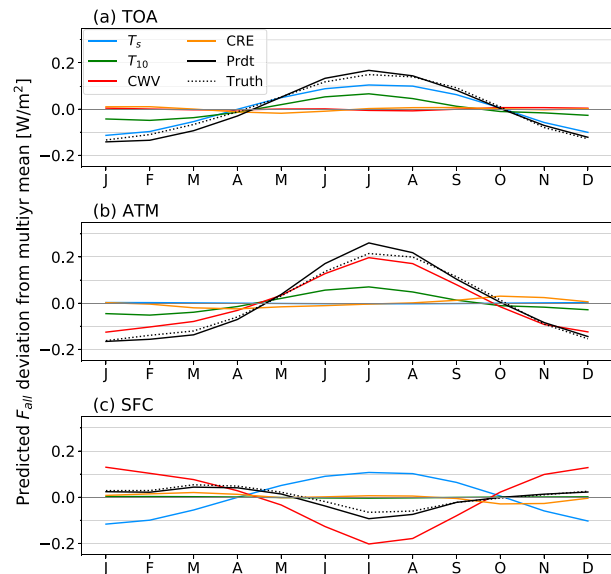


FIG. 6. The seasonality of global-mean forcing anomaly predicted by the multivariate regression model for the (a) TOA, (b) atmospheric, and (c) surface forcing. Each line identifies the partial contribution by surface temperature, T_s (blue); stratospheric temperature at 10 hPa, T_{10} (green); column water vapor, which includes the CWV and CWV^2 components (red); cloud radiative effect at the TOA or/with surface according to Eqs. (8)–(10) (orange); and the total prediction (black solid). The global-mean forcing calculated with RRTMG, the truth, is shown with the black dashed line.

Figure 7 further shows the geographical temporal forcing variability across the seasonal cycle predicted by individual predictors. The TOA, surface, and atmospheric all-sky forcing (Figs. 7a–c) all show large variability in the tropics. The atmospheric forcing, surprisingly, varies the most in Antarctica, which is mainly contributed by T_{10} variation (Fig. 7). While T_s displays obvious seasonality over Northern Hemisphere land (Figs. 7d,f), the TOA and surface forcing variability is not as large there and suggests other compensating factors, such as CWV and CRE. Interestingly, for regional TOA forcing variability, the temporal CWV variability contribution is rather small globally (Fig. 7j). The TOA forcing variability in the tropics is strongly influenced by CRE (Fig. 7m). In contrast, CWV dominates the surface and atmospheric forcing variability, especially in the Northern Hemisphere extratropics. Although CRE on the surface forcing has little seasonality in the global mean (Fig. 6c), its effect in the Arctic stands out and the magnitude is comparable to other components, such as T_s and CWV (Fig. 7o vs Figs. 7f,k).

It is important to note that we are not aiming at reducing the regression model bias by including more predictors, which could lead to overfitting. Instead, we aim to understand what makes up the forcing variability from a broad, realistic view before diving deeper into spectral aspects of the forcing. Some remaining biases, like the underestimated surface forcing in deserts and TOA forcing along west coasts, can be improved by adding surface pressure and 2-m water vapor

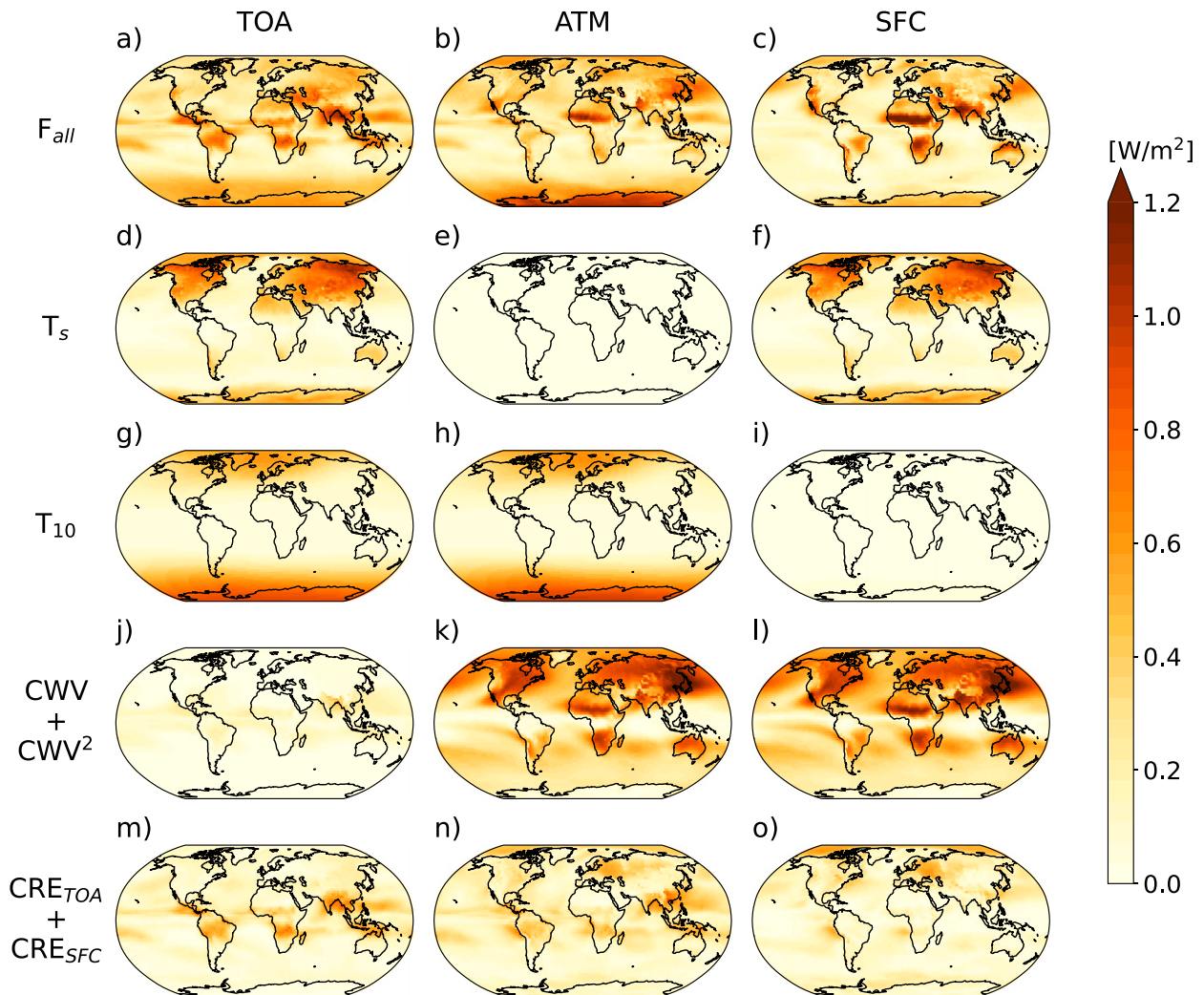


FIG. 7. As in Fig. 5, but for the temporal forcing standard deviation across the multiyear-average seasonal cycle predicted by each predictor. The forcing contributions of predictors are calculated by Eqs. (8)–(10) on a monthly basis before taking the standard deviation.

pressure as predictors. Despite being simple in form, the regression model already reveals some questions worth further exploration. For example, if the strong T_s contribution to TOA forcing is understood as the blocked surface emission, why does it end up negligible in the atmospheric forcing? Also, it is straightforward to expect that the TOA and surface forcing are mostly positive before conducting radiative transfer calculation, but why does additional CO_2 actually decrease the atmospheric forcing in high latitudes? These questions prompt a more detailed investigation into the spectral dimension of the forcing, as discussed in the next section.

5. Attributing the source of TOA forcing

a. Method

To further establish the relationship between geophysical variables and forcing variability, we examine the relationship

of TOA forcing between the surface and atmosphere from the spectral dimension via analyzing the results of RRTMG's bands. This method allows one to further link the forcing to the surface or atmospheric condition. It also bears on the two traditional views of radiative forcing, saturation of absorption bands and displacement of emission layer. A similar concept was discussed in Dufresne et al. (2020) for individual atmospheric states, and here we are offering an evaluation that covers the range of climate states across the globe.

To trace the source of radiative fluxes, we review the integrated nonscattering radiative transfer equation:

$$I(0) = I(\tau^*)e^{-\tau^*} + \int_0^{\tau^*} B(\tau)e^{-\tau} d\tau, \quad (11)$$

where I is the monochromatic radiance, and $B(\tau)$ is the thermal emission of the layer with optical depth τ associated with transmissivity $\text{Tr} = e^{-\tau}$. The optical depth is set downward positive, with zero at the TOA and τ^* at the surface. This

equation states that the radiance is the summation of transmitted emission from the source and transmitted emission by layers in between. To draw a more intuitive connection between the radiance and atmospheric profile, Eq. (11) can be alternately expressed in height coordinates as

$$I(\infty) = I(0)\text{Tr}^* + \int_0^\infty B(z)W(z)dz, \quad (12)$$

where $\text{Tr}^* \equiv e^{-\tau^*}$ is the fraction of surface emission that is transmitted to the TOA, hereby named the total transmissivity. The term $W(z)$ is the weighting function, defined as the derivative of transmission function with height:

$$W(z) = \frac{d\text{Tr}(z)}{dz} = e^{-\tau(z)} \frac{d\tau(z)}{dz}. \quad (13)$$

A typical weighting function for upward flux exhibits a bell-shaped structure that maximizes at the altitude where $\tau = 1$, meaning that the majority of TOA flux comes out from a relatively narrow range of altitudes that is referred as the “emission layer.” This allows one to define a corresponding emission temperature, but such emission temperature is highly varying spectrally as the Eq. (11) is only valid for monochromatic radiances. With instantaneous CO_2 doubling, the emission profile is unchanged in height coordinate: only $\tau(z)$ is perturbed. For instantaneous CO_2 doubling, the TOA forcing is reflected in the change of Tr^* and $W(z)$ structure, and the relative effects of Tr^* and $W(z)$ depend on the extent to which the local absorption is saturated. In a transparent, optically thin atmosphere ($\text{Tr}^* > e^{-1} \approx 0.37$), both Tr^* and $W(z)$ can change with additional CO_2 . In contrast, for regions where the absorption is saturated ($\text{Tr}^* \approx 0$), there is no room for Tr^* to decrease and therefore the change of fluxes can only come from $W(z)$ variation. We will show in what follows that CWV is highly correlated with the broadband Tr^* [defined later in Eq. (17)] in the current climate’s atmosphere, helping to quantify the role of higher specific humidity on the saturation of radiative absorption.

An immediate benefit of identifying the contributions of Tr^* and $W(z)$ to the forcing is to untangle how the temperature structure comes into play. When Tr^* is close to unity, the atmosphere is transparent. Additional CO_2 absorbs more radiance, increasing the absorptivity of the air. A weakly absorbing atmosphere can be approximated as an isothermal one, and the forcing depends on the amount of Tr^* change as well as temperature difference between the surface and atmospheric emission temperature (Dufresne et al. 2020). In contrast, in optically thick cases with large τ^* , Tr^* is close to zero, and the CO_2 forcing depends on the difference between new and old emission temperature, which is not directly related to the surface temperature. Instead, the lapse rate near the emission layer displacement would be key. With this context, Tr^* is not only a number describing the portion that the surface emission passes through the atmosphere; it also determines the extent to which the TOA forcing is connected to the surface temperature. We note that these arguments based on Tr^* and $W(z)$ are precise only for a specific spectral interval.

Given that the spectral features of Tr^* vary substantially even for an individual atmospheric state, we will later define an effective broadband Tr^* in Eq. (17) to provide intuition for the aggregate behavior.

Here, we use the optical depth in RRTMG to calculate Tr^* and combine it with the surface emission $I(0)$ to determine the transmitted surface emission that reaches the TOA [the first term on the right-hand side of Eq. (12)] and the change in TOA flux from atmospheric emission is treated as the remainder [the second term on the right-hand side of Eq. (12)]. From the difference of Eq. (12) between two CO_2 concentrations, we can separate the forcing into two components. The first forcing term, associated with total transmissivity change (ΔTr^*) and surface emission, is named as the surface contribution:

$$F_{\text{TOA_from_SFC}} = -I(0)\Delta\text{Tr}^*, \quad (14)$$

$$\Delta\text{Tr}^* \equiv \text{Tr}_{2\times\text{CO}_2}^* - \text{Tr}_{1\times\text{CO}_2}^*. \quad (15)$$

The negative sign on the right-hand side of Eq. (14) converts the forcing to downward positive, following the convention for radiative forcing. We call the rest of forcing, namely that coming from $W(z)$ change, the atmospheric contribution:

$$F_{\text{TOA_from_ATM}} = F_{\text{TOA}} - F_{\text{TOA_from_SFC}}. \quad (16)$$

We do not further decompose the $W(z)$ structure in detail. We only apply this decomposition method to the TOA forcing because $W(z)$ for the surface forcing is bottom-heavy and the effective emission height is sufficiently close to the ground that meaningful analyses require very high vertical resolution near the surface.

The analyses in this section are based on the annual-mean atmosphere under clear sky for the year 2010 to reduce the computation and storage. As the major findings do not explicitly depend on absolute location but are more related to the geophysical variables of these local conditions, we expect the results are also applicable to shorter time scales. For simplicity, the bandwise calculations are binned into three groups according to the optical properties associated with CO_2 : the center of the CO_2 absorption band ($630\text{--}700\text{ cm}^{-1}$), the wings ($500\text{--}630$ and $700\text{--}820\text{ cm}^{-1}$), and the atmospheric window ($820\text{--}1180\text{ cm}^{-1}$). The binned results are weighted by the surface emission spectrum, so the effective Tr^* in each group is calculated as

$$\text{Tr}^* = \frac{1}{I(0)} \sum_{i=0}^n I_i(0)\text{Tr}_i^*, \quad (17)$$

where the subscript i denotes the RRTMG output in the particular bands specified above. The effective broadband results are calculated over the entire longwave spectrum. The bands with weak CO_2 forcing changes (wavenumber smaller than 500 cm^{-1} or larger than 1180 cm^{-1}) constitute less than 3% of broadband forcing and are ignored here.

b. Results

Figure 8 shows the spatial pattern of total transmissivity and its change in the three grouped bands described above.

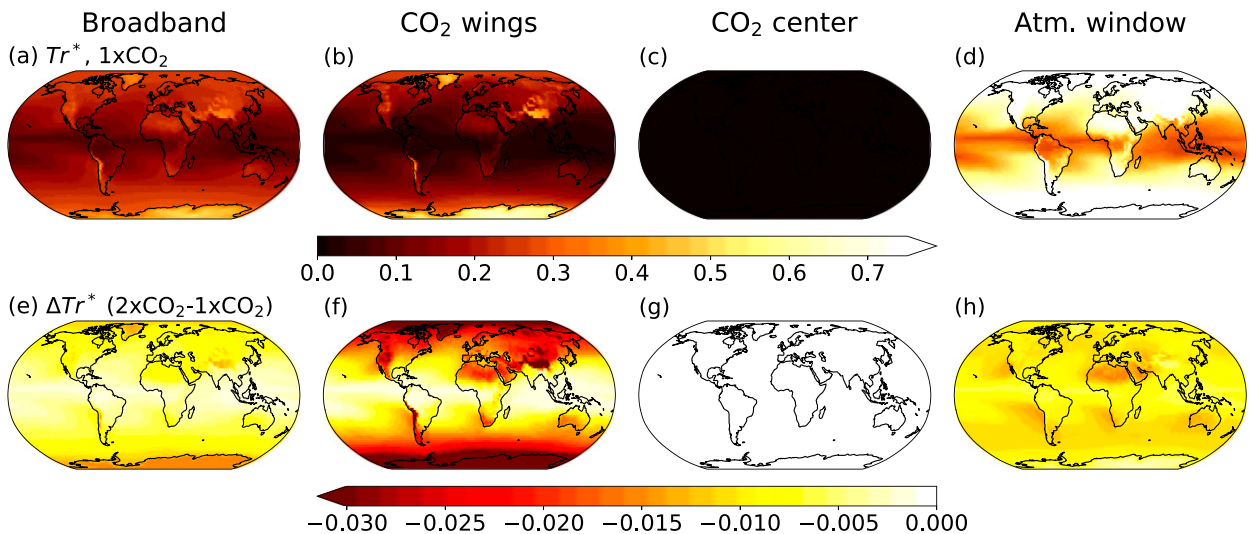


FIG. 8. (a)–(d) Total transmissivity (Tr^*) of the atmosphere and (e)–(h) total transmissivity change (ΔTr^*) with CO_2 doubling based on an annual-mean cloudless atmosphere. As the absorptivity equals to $1 - Tr^*$ in a nonscattering atmosphere, the absolute value of ΔTr^* is also the absorptivity change. In (g), all data are equal to zero. CO_2 wings include the $500\text{--}630$ and $700\text{--}820\text{ cm}^{-1}$ bands; CO_2 center refers to $630\text{--}700\text{ cm}^{-1}$; the atmospheric window contains the $820\text{--}980$, $980\text{--}1080$, and $1080\text{--}1180\text{ cm}^{-1}$ bands. The CO_2 wings, center, and atmospheric window do not add up to broadband [see (a) and (e)] as some bands (wavenumber smaller than 500 cm^{-1} or larger than 1180 cm^{-1}) are omitted.

The CO_2 band center is fully opaque, with zero Tr^* everywhere due to strong CO_2 absorption. In the wings, CO_2 absorption partly overlaps with the water vapor rotation band, particularly in $500\text{--}630\text{ cm}^{-1}$. The Tr^* distribution then reflects CWV because water vapor is the most significant greenhouse gas, with lower Tr^* in wet regions and increased Tr^* in severely dry areas like Greenland, the Tibetan Plateau, the Andes, and Antarctica. In the atmospheric window, the atmosphere is substantially less absorbing everywhere, yet it still displays a pattern resembling the climatology of CWV.

The change of total transmissivity (ΔTr^*) with CO_2 doubling does not alter the global pattern of Tr^* , as ΔTr^* is an order of magnitude smaller than Tr^* (Figs. 8e–h). The Tr^* in the CO_2 band center is unchanged by increased CO_2 as the absorption is already saturated at the control concentration. For the CO_2 wings, ΔTr^* exhibits a pattern that is the opposite of the climatological Tr^* , with more ΔTr^* decrease in the polar regions and deserts, where the column is relatively transparent. This pattern also underlies the water vapor masking effect—that the water vapor and CO_2 overlapping at an absorbing band inhibits absorption by additional CO_2 . Although considerably weaker, the masking effect also presents in the atmospheric window.

Figures 9a–d compare Tr^* , ΔTr^* , and their connections with CWV in further detail. Aside from the CO_2 band center, it is apparent that the Tr^* increases in drier columns, accompanied with larger magnitude of ΔTr^* . Interestingly, ΔTr^* appears linearly related to Tr^* , except for the driest locations. The driest locations, where CWV is less than $\approx 10\text{ kg m}^{-2}$, correspond to regions such as Antarctica, Greenland, and the Tibetan Plateau, where the terrain is elevated and the surface temperature is extremely low.

The pattern of surface and atmospheric contribution to the TOA forcing is shown in Fig. 10. The forcing contribution from the surface stems from nonzero ΔTr^* in the CO_2 wings and atmospheric window (Figs. 10a,b,d). The water vapor masking effect is clear in the CO_2 wings, as the forcing is largely reduced in the tropics. Since no surface emission can penetrate the optically thick atmosphere in the CO_2 absorption band center, the surface does not contribute to the TOA forcing in this band and all of the forcing comes from the atmosphere (Figs. 10c,g). The sign of surface contribution is nonnegative everywhere, while the atmospheric contribution can be either positive or negative. In the CO_2 band center, the emission layer is located within the stratosphere, and when the emission layer is raised by extra CO_2 , the negative lapse rate there leads to an increase in longwave cooling and therefore the atmosphere itself negatively contributes to the forcing (Fig. 10g; Huang et al. 2016; Jeevanjee et al. 2021). Similar reasoning is also applicable to positive atmospheric contribution in CO_2 wings. The only difference is that the emission layer is moved inside the troposphere, and the positive lapse rate there means that the added CO_2 decreases the emission temperature (Fig. 10f).

A counterintuitive feature is that the atmosphere can negatively contribute to the TOA forcing outside of CO_2 absorption band. This is seen in the high latitudes of CO_2 wings and everywhere in the window band (Figs. 10f,h). These regions are optically thin (Figs. 8b,d) and are not expected to have an emission layer within the stratosphere, which would allow the negative stratospheric lapse rate argument to be invoked. Instead, in these weak absorbing regions with $Tr^* > e^{-1}$, $W(z)$ appears to have a bottom-heavy profile. As ΔTr^* is much smaller than Tr^* , the emission layer remains at the bottom of

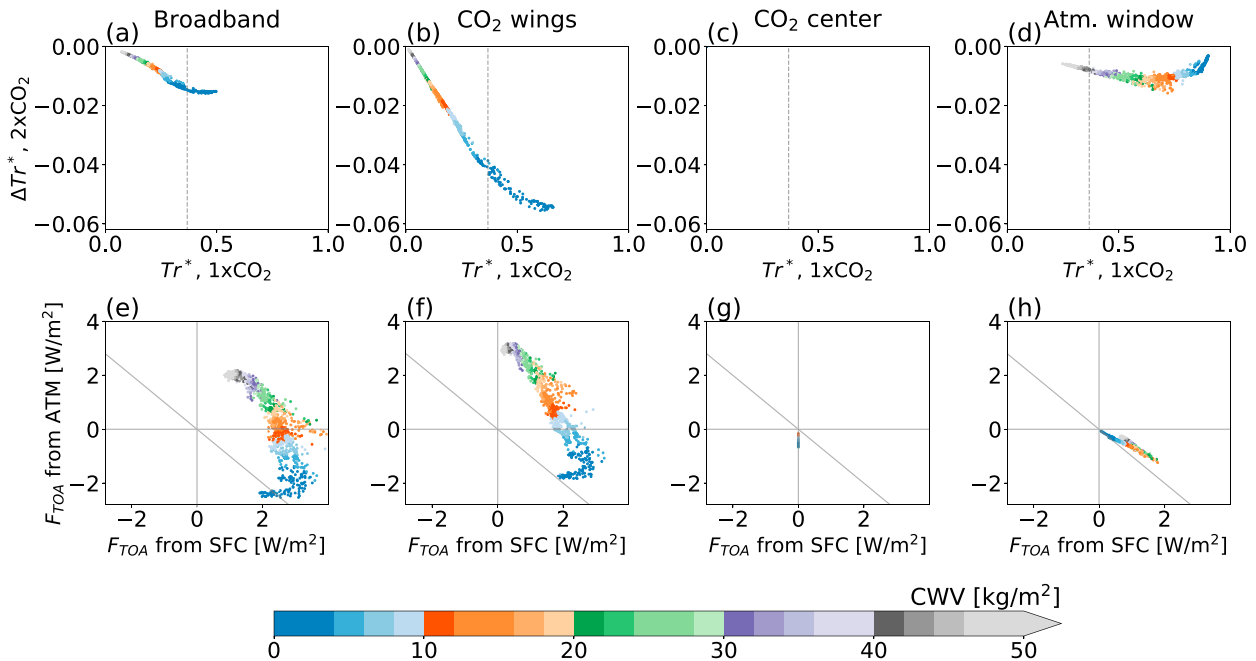


FIG. 9. (a)–(d) The change of total transmissivity (ΔTr^*) from CO_2 doubling vs total transmissivity (Tr^*) of the control CO_2 . Each dot represents a single grid in the global map and is color coded by column water vapor ($kg\ m^{-2}$). The gray dashed line denotes $Tr^* = e^{-1} \approx 0.37$. When $Tr^* < e^{-1}$, $W(z)$ is bell-shaped with an emission layer $\tau = 1$ within the atmosphere. Otherwise, $W(z)$ would be bottom-heavy with $Tr^* > e^{-1}$. The emission layer would be in the lowermost atmosphere but is not identical to traditionally defined $\tau = 1$ layer. (e)–(h) The TOA forcing originating from the atmosphere vs the TOA forcing originating from the surface, or, equivalently, the TOA forcing contributed by the emission layer change vs absorption change [Eqs. (16) and (14), respectively]. The gray slanted line is $x + y = 0$ and the distance from this line equals the magnitude of TOA forcing. No data are shown in (c) because the absorption in CO_2 band center is fully saturated and therefore Tr^* and ΔTr^* are both equal to zero. The grouping of bands is the same as in Fig. 8.

the column and does not shift with additional CO_2 . The additional CO_2 reduces transmissivity (Tr^*), while absorptivity ($1 - Tr^*$) increases. As a result, the emissivity is increased in accordance with Kirchhoff's law of radiation. With increased emissivity, there is more emission from the atmosphere to space. Although such a transparent atmosphere reduces the forcing, it does not override the surface contribution and the forcing in the particular band is still positive (Figs. 9f,h and 10j,l). As shown in Figs. 9e–h, the bandwise TOA forcing is always positive in CO_2 wings and atmospheric window yet is very close to zero in the Antarctic, where CWV is less than $5\ kg\ m^{-2}$. Figures 9e–h also show that the negative broadband TOA forcing over Antarctica results from relatively strong cooling in the CO_2 band center (Figs. 9g and 10k). This feature is captured by T_{10} variability in the regression model [Eq. (2)].

Figure 11 presents the relationship between the surface and atmospheric forcing and the decomposed TOA forcing components. The TOA forcing originating from the atmosphere matches the atmospheric forcing well, and similarly the TOA forcing coming from the surface is very close to the surface forcing in most bands. The only inconsistency occurs in the CO_2 band center, which is particularly opaque. Here, the TOA and surface forcing are independent since the emission layer is very high for outgoing longwave at the TOA, while being very close to the surface for the downward surface flux, and the surface emission barely reaches to the TOA.

By physically decomposing the source of TOA forcing based on spectroscopy, the choice of predictors in the regression models is justified and further reveals the intrinsic difference between the surface and atmospheric forcing. For the surface forcing, CWV explains a large percentage of surface forcing pattern as it is highly correlated to ΔTr^* (Figs. 9a–d), but the surface temperature is still essential for satisfying surface forcing predictions since the surface emission can only be represented by T_s (Fig. 5c). In contrast, the atmospheric forcing can be qualitatively captured by CWV alone, as CWV greatly shapes the transmissivity pattern.

6. Conclusions and discussion

In this study, we thoroughly investigated the instantaneous forcing of CO_2 doubling at the TOA, at the surface, and within the atmosphere. We demonstrated how the forcing patterns exhibit significant seasonal and geographical variability. The TOA forcing has clear meridional structure in the multiyear climatology, whereas the surface forcing is more variable zonally and is heavily influenced by humidity. Overall, the atmospheric forcing in the deep tropics is quite smooth, yet it exhibits a noticeable land–sea contrast with the opposite sign of forcing in the subtropics and midlatitudes. Over the subtropics and midlatitudes, the atmospheric forcing is generally negative over land, implying that increased CO_2 enhances atmospheric

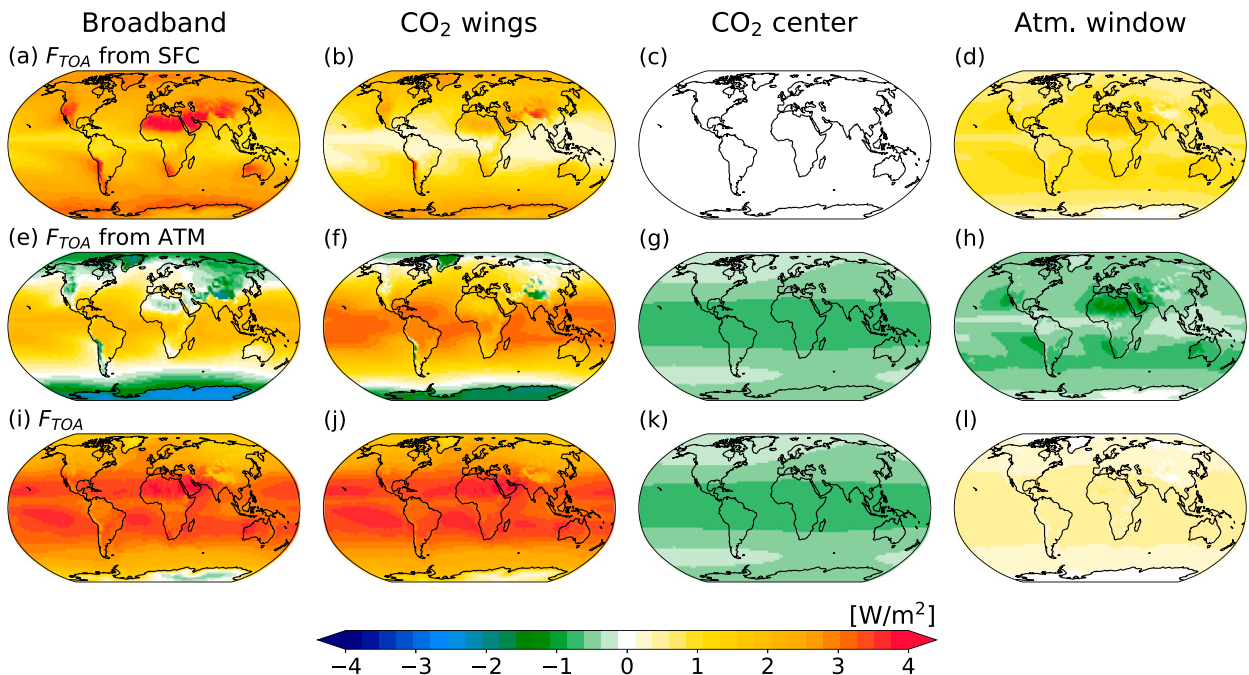


FIG. 10. As in Fig. 8, but for (a)–(d) the spatial pattern of TOA forcing originating from the surface [Eq. (14)], (e)–(h) TOA forcing originating from the atmosphere [Eq. (16)], and (i)–(l) the total TOA forcing in the bands (shown from left to right in columns) indicated by the titles. The first and second rows sum to the bottom row. All data in (c) are equal to zero.

longwave cooling, which has not been highlighted in the past. We note that this feature is not robust in climate models, some of which have positive CO_2 atmospheric forcing over land [see Fig. 1e herein and Figs. 7f and S1 of Chadwick et al. (2019)]. The strong surface forcing may lead to rapid warming over land regions, so the radiative flux changes over land will depart from the instantaneous forcing even on short time scales. In terms of temporal variation, the forcing is rather steady between years, but displays significant seasonality. The seasonality of global-mean surface forcing goes in the opposite direction to the global-mean TOA forcing. The TOA forcing maximizes in Northern Hemisphere summer, which is also the season with the minimum surface forcing. Therefore, the atmospheric forcing in boreal summer is about twice as strong as that in boreal winter, implying a nonnegligible seasonality on the global-mean atmospheric energy budget.

Then we used multivariate regression models, a convenient method to estimate the forcing pattern with the information of just surface temperature, stratospheric temperature, column water vapor, and cloud radiative effect. We showed that the TOA forcing is mostly determined by the surface and stratospheric temperatures, with some modifications made by the cloud and water vapor masking effects. In contrast, water vapor explains the majority of the variation in the surface forcing, although one must also take surface temperature into account. Interestingly, we found that the surface temperature's contribution to the TOA and surface forcing is nearly identical, and the surface temperature has a small-magnitude contribution for the atmospheric forcing regression model. Also, the regression model does not contain information

beyond the surface and stratospheric temperature but can account for the high-latitude forcing, suggesting that it is worthwhile to perform a detailed investigation of the role of temperature inversions on the forcing. These models successfully capture the majority of the forcing inhomogeneity and seasonality, despite their simplicity. Since past radiative transfer model intercomparisons documented that the inaccuracy of radiative forcing due to radiation parameterizations is generally smaller compared to the forcing value variations caused by state dependence on temperature, water vapor, and clouds (Ellingson et al. 1991; Collins et al. 2006; Pincus et al. 2020), the regression coefficients determined in this study provide a measure of cloud and water vapor effects. The regression models can therefore be used in future works to quantify the intermodel forcing difference caused by climate fields.

Given the logarithmic dependence of radiative forcing on gas concentration, it is possible to extend the regression models based on CO_2 doubling to obtain the predictive equations for radiative forcing of CO_2 quadrupling by multiplying Eqs. (2)–(10) by a factor of 2. The logarithmic dependence of instantaneous TOA forcing has been shown to hold fairly well in Huang and Bani Shahabadi (2014). However, more bias might arise for the surface forcing because a large portion of surface forcing is related to the transmissivity change (Figs. 11e–h), and the logarithmic dependence expected from the emission layer displacement may not apply.

Furthermore, water vapor plays a crucial role in determining the forcing pattern, particularly for the atmospheric forcing. This reflects the relative importance of absorption band saturation and emission layer displacement in different Tr^* regimes.

cooling. How much of the respective forcing may give rise to large-scale circulation changes as opposed to local rebalancing via column adjustments and land surface temperature change warrants future investigation.

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Data availability statement. The ERA5 and ERA5-land datasets can be accessed through the ECMWF website (<https://cds.climate.copernicus.eu/>). The RRTMG code can be downloaded at http://rtweb.aer.com/rrtm_frame.html. The forcing dataset and scripts for figures are available at <https://doi.org/10.5281/zenodo.7076271>.

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