Transient Extratropical Response to Solar Ultraviolet Radiation in the Northern Hemisphere Winter

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ABSTRACT: An intermediate complexity general circulation model is used to investigate the transient response of the NH winter stratosphere to modulated ultraviolet (UV) radiation by imposing a stepwise, deliberately exaggerated UV perturbation and analyzing the lagged response. Enhanced UV radiation is accompanied by an immediate warming of the tropical upper stratosphere. The warming then spreads into the winter subtropics due to an accelerated Brewer–Dobson circulation in the tropical upper stratosphere. The poleward meridional velocity in the subtropics leads to an increase in zonal wind in midlatitudes between 20° and 50°N due to Coriolis torque. The increase in midlatitude zonal wind is accompanied by a dipole in Eliassen–Palm flux convergence, with decreased convergence near the winter pole and increased convergence in midlatitudes (where winds are strengthening due to the Coriolis torque); this dipole subsequently extends the anomalous westerlies to subpolar latitudes within the first 10 days. The initial radiatively driven acceleration of the Brewer–Dobson circulation due to enhanced shortwave absorption is replaced in the subpolar winter stratosphere by a wave-driven deceleration of the Brewer–Dobson circulation, and after a month the wave-driven deceleration of the Brewer–Dobson circulation encompasses most of the winter stratosphere. Approximately a month after UV is first modified, a significant poleward jet shift is evident in the troposphere. The results of this study may have implications for the observed stratospheric and tropospheric responses to solar variability associated with the 27-day solar rotation period, and also to solar variability on longer time scales.

KEYWORDS: Stratospheric circulation; Stratosphere-troposphere coupling; Solar cycle

1. Introduction

The sun is the source of energy for Earth's climate system. Solar output varies on time scales ranging from the subseasonal to the millennial, and on each time scale an atmospheric response can be detected (Rind 2002) although it may be small relative to that of other climate forcings (Feulner and Rahmstorf 2010; Schurer et al. 2014). One of these sources for solar variability is the 11-yr solar cycle. The solar cycle is manifested by changes in the density of sunspots on the sun's surface. While the sunspot itself is darker (colder) than the photosphere's background, it is surrounded by a much brighter (warmer) boundary of faculae. On average, the faculae brightening dominates, and the solar cycle produces changes of approximately 1 W m^{-2} in total solar irradiance (TSI) (Foukal et al. 2006), mainly in the ultraviolet (UV) components of the solar spectrum. While the solar cycle's contribution to TSI is on the order of 0.1%, changes in the UV near 200 nm can approach 8% (Ermolli et al. 2013; see also Fig. 3 of Coddington et al. 2016). Evidence of past solar activity suggests that the irradiance increased from the Maunder Minimum to the present by about 26.6% in the Schumann–Runge bands and 10.9% in the Herzberg continuum (Shapiro et al. 2011), although other estimates suggest a smaller

change (Lean 2000) and the change in total solar irradiance estimated by Shapiro et al. (2011) may be too large (Coddington et al. 2016; Egorova et al. 2018).

Intensification of the solar UV radiation leads to up to a $\sim 2\text{-K}$ warming in the tropical upper stratosphere during solar maximum years in the zonal mean (Crooks and Gray 2005; Shibata and Deushi 2008; Frame and Gray 2010; Sukhodolov et al. 2017) although the magnitude differs among reanalysis products (Mitchell et al. 2015a). This warming of the upper tropical stratosphere is due to increased absorption by ozone of radiation with wavelengths between 240 and 320 nm (Haigh 1994; Gray et al. 2010). Ozone concentrations are also increased during solar maximum due to the enhanced radiation at 200 nm where oxygen dissociation and ozone production occur (Gray et al. 2010) and this enhancement in ozone in turn enhances the warming in the tropical upper stratosphere (Haigh 1994; Shindell et al. 1999; Swartz et al. 2012; Bednarz et al. 2019). The meridional temperature gradient in the winter subtropical upper stratosphere is therefore expected to become more negative during the solar maximum, and hence anomalously strong westerlies should be present in the subtropics as well. Such changes in subtropical wind have been detected in response to solar variability both on the 27-day rotation period (Hood 2004; Garfinkel et al. 2015) and on longer time scales (Matthes et al. 2006; Mitchell et al. 2015a,b; Lu et al. 2017a,b). There is abundant evidence that this subtropical acceleration subsequently extends farther poleward in modeling experiments (Shindell et al. 1999; Tourpali et al. 2003; Egorova et al. 2004, to name a few) as reviewed by Gray et al. (2010), although the magnitude of this effect in reanalysis data is weak (e.g., Fig. 9 of Mitchell et al. 2015a). After the vortex is strengthened during the solar maximum, the effect

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appears to propagate downward by eddy mean–flow interactions and descend toward the troposphere (Matthes et al. 2006; Ineson et al. 2011; Chiodo et al. 2012; Lu et al. 2017a,b), where it induces variations in the tropospheric eddy-driven jets and modifies surface temperatures and precipitation over the North Atlantic and Europe (Haigh 1996; Ineson et al. 2011; Gray et al. 2013), although the observed surface signal may also be associated with unforced variability (Chiodo et al. 2019). This mechanism is referred to as the “stratospheric” or “top-down” pathway for the solar influence on surface climate (Mitchell et al. 2015b).

The tropical and extratropical stratospheres are connected via the Brewer–Dobson circulation (BDC; Dobson et al. 1929; Brewer 1949; Dobson 1956; Butchart 2014). The BDC includes air rising through the tropical tropopause and into the tropical stratosphere, and downward motion in the winter stratosphere. The BDC is driven by breaking and/or dissipation of tropospheric planetary (Rossby) and gravity waves, and also by diabatic heating, which is required for cross-isentropic motion (Butchart 2014). As the solar cycle modulates diabatic heating directly and also alters the background state encountered by upward propagating waves, it is reasonable to expect changes in the BDC (Shibata and Kodera 2005). Specifically, solar maximum has been linked to a weakened BDC (Kodera and Kuroda 2002) and a secondary maximum in temperature in the lower stratosphere (Hood and Soukharev 2012), although models have trouble simulating this effect (Mitchell et al. 2015b) and the observed secondary maximum may be due to aliasing from volcanoes or sea surface temperature variability (Marsh and Garcia 2007; Chiodo et al. 2014; Kuchar et al. 2017).

This work is motivated by the following uncertainties in the extratropical response to solar variability. First, the time scale over which the stratospheric pathway develops has received relatively little attention, but this time scale is crucially important for extratropical impacts from the 27-day solar rotation. If the time scale for the stratospheric pathway is longer than the rotation period, then there is no expectation for the mechanism to operate on subseasonal time scales, and therefore the 27-day solar rotation period cannot be used to enhance subseasonal forecasting in the troposphere via this top-down pathway (although other pathways may still exist; Hood 2018). On the other hand, if the time scale is shorter than the 27-day rotation period, then one should expect a surface response assuming eddy–mean flow interaction can propagate the signal downward.

Second, the response of the BDC to solar variability is still unclear: enhanced solar flux directly leads to enhanced shortwave heating, and thus one might naively expect accelerated cross-isentropic transport and a faster BDC in the tropics just as, for example, aerosol heating has been shown to lead to a speeding up of the tropical upwelling (Robock 2000; Aquila et al. 2014; Garfinkel et al. 2017) although we acknowledge the potential for differences in the response to heating in the lower versus upper stratosphere. However, previous observational work (e.g., Kodera and Kuroda 2002) suggests that the BDC is weakened by solar maximum.

More generally, the goal of this work is to understand the time development of the impacts of solar ultraviolet radiation.

Many previous studies of the top-down pathway used long runs of climate models with observed slowly varying or fixed solar variations (e.g., Matthes et al. 2006). The limitation of such an approach is that internal stratospheric feedbacks are faster than the characteristic time scales of observed UV variability, and hence the initial mechanisms whereby UV variability affects the stratosphere is obscured by feedbacks operating within the winter stratosphere that are generic to any forcing (e.g., Watson and Gray 2015). Here we adopt a different approach and analyze idealized model experiments in which we switch on anomalous solar ultraviolet radiation, and hence we can focus on the transient development of the extratropical stratospheric response to solar variations. Our particular focus is on how quickly UV variations leads to changes in the stratospheric polar vortex at latitudes where there is little solar absorption to begin with, and also how enhanced UV radiation can lead to a weakened BDC even as shortwave heating is enhanced. Such a transient approach has been used, for example, to understand the mechanisms for shifts in the tropospheric circulation in response to enhanced greenhouse gases (Chemke and Polvani 2019) or to the quasi-biennial oscillation (Garfinkel and Hartmann 2011; Garfinkel et al. 2012).

After introducing the model in section 2, we show that anomalous solar ultraviolet radiation leads to a time-dependent response in which wave fluxes play a crucial role in extending the initial subtropical upper stratospheric perturbation toward polar latitudes in the lower stratosphere, and subsequently to the surface. These wave fluxes also lead to nonmonotonic changes in the BDC: the initial radiatively driven acceleration of the BDC for enhanced UV radiation is replaced through most of the stratosphere by a dynamically driven deceleration. Implications for the surface impacts from observed solar variability on short (e.g., the 27-day solar rotation period) and long time scales are discussed in section 4, and the results are summarized in section 5.

2. Methods

We use the model of an idealized moist atmosphere (MiMA) introduced by Jucker and Gerber (2017) with the extensions introduced by Garfinkel et al. (2020). Full details of the model are described in detail in Jucker and Gerber (2017), and here we briefly introduce its key aspects. This model is a primitive equation model with moisture (and latent heat release), a mixed layer ocean, Betts–Miller convection (Betts 1986; Betts and Miller 1986), and a boundary layer scheme based on Monin–Obukhov similarity theory [following Frierson et al. (2006, 2007) and Merlis et al. (2013)]. Neither a sponge-layer nor Rayleigh damping scheme is utilized; instead, the gravity wave scheme of Alexander and Dunkerton (1999) is used to represent gravity wave momentum deposition following Cohen et al. (2013). The gravity wave scheme is also modified to ensure that any gravity wave momentum fluxes that do reach close to the model lid are deposited in the levels above 0.85 hPa so as to avoid possible sponge-layer feedbacks and spurious meridional circulations (Shepherd and Shaw 2004; Shaw and Shepherd 2007). Realistic topography, land–sea contrast, and horizontal ocean heat fluxes are imposed as in Garfinkel et al. (2020) in order to represent stationary waves as well as comprehensive

FIG. 1. Shortwave heating rate in the first 2 months after branching in MiMA with (a) enhanced UV, (b) default UV, and (c) reduced UV. The contour interval is 1.75 K day^{-1} . Note that the range of the abscissa differs in this figure in order to show the heating rates in the summer hemisphere.

climate models, and the stratospheric sudden warming (SSW) frequency is 3.8 instances per decade at T42 resolution and 5.5 per decade at T85 resolution. The internal stratospheric time scales associated with SSWs in this model as compared to reanalysis were examined by White et al. (2020) and found to be realistic (their Fig. 2a).

MiMA incorporates the Rapid Radiative Transfer Model (RRTM) radiation scheme (Mlawer et al. 1997; Iacono et al. 2000), unlike some more idealized moist aquaplanet models that use the gray-radiation scheme of Frierson et al. (2006) and hence cannot resolve the interaction of shortwave radiation with ozone. With this radiation scheme, we are able to incorporate the radiative impacts of ozone and water vapor into the model. To isolate the impact of solar UV variations on the stratosphere, we magnify the anomaly in solar UV. Specifically, the default source of shortwave radiation in RRTM is multiplied by $4/3$ for the enhanced UV integration (hereafter enhanced UV), and by $2/3$ for the reduced UV integration (hereafter reduced UV), in the bands of RRTM that correspond to UV (200–340 nm). Note that no changes are made elsewhere in the spectrum. See the appendix for the detailed implementation of this within MiMA. These UV changes lead to a $\sim 1\%$ variation in TSI, and while this greatly overestimates the impacts of the contemporary solar cycle, during the Maunder Minimum, solar irradiance may have been 25% lower than the present near 200 nm (Shapiro et al. 2011). Furthermore, photolysis effects from rising ozone concentration in response to solar maximum provide a positive feedback to the radiative effects (Haigh 1994; Shindell et al. 1999; Swartz et al. 2012; Bednarz et al. 2019), and part of the exaggerated solar forcing can be thought of as representing this ozone feedback, which is not included in the fixed ozone concentrations used here. The net effect on shortwave heating rates averaged over January and February is shown in Fig. 1, and in the summer hemisphere the heating rates are twice as large for enhanced UV as compared to reduced UV. Note that the difference in heating rate between reduced and normal UV

slightly exceeds that between enhanced and normal UV by $\sim 4\%$. These shortwave heating rates (and also longwave heating rates shown later in this paper) are taken from the model's radiation scheme.

To better understand the stratospheric pathway, we focus on the temporal evolution of anomalies in switch-on experiments in which the incoming solar radiation is changed abruptly. Specifically, we start with a long control run of MiMA with normal solar conditions and a full seasonal cycle, and branch off every 1 January with 29 simulations each of enhanced and reduced UV radiation. The perturbation in the solar forcing relative to the control solar forcing is held fixed over the duration of these switch-on experiments, and our focus here is on the first 50 days after branching (i.e., through mid-February). By day 50, anomalies have reached statistical steady state in both the stratosphere and troposphere. Switch-on experiments of the response to solar variability were performed by Simpson et al. (2009), but their focus was on the tropospheric response to the lower stratospheric tropical secondary maximum imposed in a model that did not represent radiation in a physically consistent manner. We have also created an ensemble branching from 1 November of the control run, and results from this sensitivity ensemble are shown in the online supplemental material.

Anomalies are computed by comparing the enhanced UV and reduced UV branch runs to the control run with default UV radiation. Statistical significance is computed using a two-tailed Student's t difference of means test between the enhanced and reduced UV branch runs. Note that statistical significance can also be computed by comparing the enhanced UV-normal response to the reduced UV-normal response, and for such a comparison statistical significance is more extensive and all features discussed in this paper are significant. Furthermore, all anomalies shown here are forced by the altered UV because we are comparing matched pairs of simulations (e.g., O'Sullivan and Young 1992; Garfinkel et al. 2012). Therefore, even anomalies not marked as statistically significant by the Student's t test are a direct response to the altered UV.

The TEM zonal momentum and thermodynamic equations are used to diagnose how the stratosphere responds to altered UV [appendix of chapter 12 in Vallis (2006)]. The divergence of the Eliassen–Palm flux (EPF) $\mathbf{F}$

$$\nabla \cdot \mathbf{F} = \frac{1}{a \cos \varphi} \frac{\partial}{\partial \varphi} (F^{(\varphi)} \cos \varphi) + \frac{\partial F^{(z)}}{\partial z} \quad (1)$$

drives changes in zonal wind in the zonal-mean zonal momentum budget:

$$\frac{\partial \bar{u}}{\partial t} = \bar{v}^* \left[f - \frac{(\bar{u} \cos \varphi)_{\varphi}}{a \cos \varphi} \right] - \bar{w}^* \frac{\partial \bar{u}}{\partial z} + \frac{\nabla \cdot \mathbf{F}}{\rho_0 a \cos \varphi} + \bar{X}. \quad (2)$$

In these equations, z is the log-pressure height, v and w are the meridional and vertical components of the wind, θ is the potential temperature, and a , f , and ρ_0 are Earth's radius, the Coriolis parameter, and the background density profile, respectively. Overbars and primes represent zonal averages and the deviations therefrom, respectively. The meridional and vertical components of the EP flux in spherical coordinates are

$$F^{(\varphi)} = a \rho_0 \cos \varphi \left(\bar{u}_z \frac{\overline{v'\theta'}}{\bar{\theta}_z} - \overline{u'v'} \right), \quad (3a)$$

$$F^{(z)} = a \rho_0 \cos \varphi \left\{ \left[f - \frac{(\bar{u} \cos \varphi)_{\varphi}}{a \cos \varphi} \right] \frac{\overline{v'\theta'}}{\bar{\theta}_z} \right\}. \quad (3b)$$

Note that the $\overline{u'w'}$ term is not included here.

The TEM thermodynamic equation

$$\frac{\partial \bar{T}}{\partial t} + \frac{\bar{T}_{\varphi}}{a} \bar{v}^* + \frac{N^2 H}{R} \bar{w}^* = \frac{J}{C_p} \quad (4)$$

is used to understand how zonal mean temperatures change in response to altered incoming UV [following Eq. (10.18) in Holton and Hakim (2013)].

For simplicity we equate the BDC with the TEM residual circulation; however, we recognize that mass transport is also affected by mixing processes. The TEM residual circulation is computed by vertically integrating $\bar{v}^*$ for the streamfunction:

$$\Psi^* = \int_z^{\infty} \rho_0 \bar{v}^* \cos \varphi dz$$

following Eq. (10) of White et al. (2020).

3. Results

We first consider the linearity of the impact from enhanced UV as compared to reduced UV. We then examine the development of the stratospheric pathway from the tropical upper stratosphere to the midlatitude upper stratosphere, and subsequently to the subpolar region and down to the tropopause. To explain this evolution, we analyze the thermodynamic and zonal momentum transformed Eulerian mean (TEM) budgets and diagnose the mechanism whereby the initial equatorial heating anomaly impacts the rest of the stratosphere. We conclude by considering changes in stratospheric

sudden warming frequency and the tropospheric impacts. As shown in White et al. (2020), the internal stratospheric time scales in this model are realistic (see their Fig. 2a), and hence this model can be used to clarify the relevant time scales in response to an external forcing.

We first consider the linearity of the response to the simulated solar forcing by separately considering differences between enhanced UV and normal conditions (Fig. 2, left column) and normal and reduced UV conditions (Fig. 2, right column), averaged over the first 60 days. A 33% change in incoming UV leads to zonal-mean temperature anomalies exceeding 15 K, and zonal-mean zonal wind anomalies approaching 15 m s⁻¹. The subpolar zonal wind anomaly is related to the subtropical temperature anomaly through thermal wind balance. The difference between enhanced UV and normal conditions is similar to that between normal and reduced UV conditions although the difference is slightly larger for reduced UV consistent with the slightly larger perturbation to the shortwave heating rate (section 2), and hence the response is nearly linear.

Figure 3a considers the transient evolution of the upper-stratospheric tropical temperature response to enhanced UV (red) and reduced UV (blue) separately, and indicates that the response to reduced UV is slightly larger throughout the duration of our integrations (deduced from the distance from the zero line), again consistent with the larger perturbation to the shortwave heating rate, although deviations from linearity are small. The extratropical transient wind responses in the upper stratosphere (Fig. 3b) are also nearly linear, although slightly stronger for reduced UV as before. In the lower stratosphere, enhanced UV flux leads to a slightly stronger response in subtropical zonal wind (Fig. 3c) while reduced UV flux leads to a slightly stronger response farther poleward (Fig. 3d), although deviations from linearity are not statistically significant. The difference in winds between the subtropics and midlatitudes (Fig. 3d minus Fig. 3c) is similar regardless of whether UV flux is reduced or enhanced. Responses are generally similar if we focus on the ensemble of branch runs initialized on 1 November (Figs. S2 and S3 in the online supplemental material). In the rest of this work we compare the enhanced UV branch runs to the reduced UV branch runs in order to enhance the signal-to-noise ratio. We discuss the solar impact in terms of enhanced UV minus reduced UV, and while the quantitative details differ depending on the sign of the perturbation, each argument is equally valid for the reverse case of reduced UV.

The temperature response in the 60-day average in Fig. 2 extends to latitudes where little solar radiation is received and differences in the shortwave heating rates among the experiments are small (Fig. 1), already indicating that dynamical feedbacks have operated within the stratosphere during the first 60 days.

How does this extratropical response develop? The temperature response is shown in Fig. 4: enhanced UV leads to warming in the first two days mostly in the same region where shortwave heating is increased, and the initial rise in temperature in the tropics occurs within 10 days (Fig. 3a) consistent with the fast radiative time scales in this region. During days 3–9 the warming expands northward and downward with time

FIG. 2. (top) Zonal-mean temperatures and (bottom) zonal winds, for (right) reduced and (left) enhanced UV averaged over the first 60 days after branching. Black dots signify statistical significance at the 5% level. The contour interval is 2 K or 1.5 m s⁻¹.

(Fig. 4b), and by days 10–19 the temperature response has plateaued in the tropical upper stratosphere (Fig. 3a; the TEM thermodynamic balance will be used to explain the plateauing later) although the region with warming grows and encompasses

most of the stratosphere at later lags (Figs. 4c,d). The midlatitude upper-stratospheric zonal wind anomaly already appears in the first two days (Fig. 5a), and it strengthens and propagates downward alongside the temperature gradient over the

FIG. 3. Temporal evolution of the difference between the control run and the two branch runs in (a) temperature at 2 hPa and 2°S–2°N, (b) zonal wind at 2 hPa and 55°–65°N, (c) zonal wind at 51 hPa and 25°–35°N, and (d) zonal wind at 51 hPa and 55°–65°N. Vertical lines indicate the 95% confidence intervals on the modeled response.

FIG. 4. Difference in Northern Hemisphere zonal mean temperature between enhanced UV and reduced UV branch runs in (a) days 1 and 2, (b) days 3–9, (c) days 10–19, and (d) days 20–29 after branching. Dots indicate where a null hypothesis of no difference can be rejected at the 5% confidence level. The contour interval is 2.5 K. The black contour shows the 4 and 10 m s^{-1} isotachs of zonal wind.

following month (Figs. 5b–d and 3b), before plateauing after around a month (Fig. 3b). A strengthening of zonal wind in the subpolar lower stratosphere and a weakening of zonal wind in the subtropical lower stratosphere develop over the first month after branching (Fig. 5d), and also plateau after around a month (Figs. 3c,d).

In contrast to the temperature and zonal wind responses, which are monotonic in time, the TEM streamfunction response is nonmonotonic (Fig. 6). At first, a positive streamfunction anomaly develops throughout the tropical and subtropical stratosphere, with anomalous rising motion at the equator and sinking in midlatitudes (i.e., enhancing the climatological BDC at these latitudes; Fig. 6a). This strengthening can be thought of as a direct response to enhanced diabatic

FIG. 5. As in Fig. 4, but for zonal wind. The contour interval for colored contours is 2.5 m s^{-1} , and the 0.5 m s^{-1} contour is in black and the -0.5 m s^{-1} contour is blue.

heating at these latitudes when solar radiation is increased, as will be discussed in more detail later. This positive streamfunction anomaly is later replaced throughout most of the stratosphere with a weakening of the BDC (the tropical upper stratosphere is the only exception). This weakening begins first at polar latitudes in the lower stratosphere (Fig. 6b), before expanding to the polar upper stratosphere and tropical lower stratosphere (Figs. 6c,d). This weakened BDC is associated with weak cooling over the pole in response to enhanced UV (Fig. 4d and more clearly in Fig. S4d). The temperature, zonal wind, and streamfunction responses are generally similar if we focus on the ensemble of branch runs initialized on 1 November (Figs. S4–S6), although the time scale over which the anomalies develop is longer than for the 1 January ensemble.

What drives the nonmonotonic behavior of the BDC? This nonmonotonic behavior can be explained, in a diagnostic sense, with the TEM thermodynamic and zonal momentum balances, as we now demonstrate.

stream function anomaly, enhanced–reduced UV

FIG. 6. As in Fig. 4, but for the TEM streamfunction (a metric of the BDC). The contour interval is $2 \text{ kg m}^{-1} \text{ s}^{-1}$. Note that the ranges shown for both the abscissa and ordinate differ from those in other figures in order to focus on the stratosphere.

We show the TEM thermodynamic balance as displayed in Eq. (4) in Fig. 7. Enhanced UV leads to enhanced shortwave heating in the deep tropics, with net radiative heating peaking at more than 5 K per day averaged over the first two days (Fig. 7c). This leads to local warming and therefore enhanced longwave cooling [not shown explicitly, but see the difference between Figs. 7c and 7i; as in Haigh (1994)]. The longwave feedback does not balance the shortwave tropical heating fully even at later lags (Figs. 7i,l), and the residual radiative heating is balanced by anomalous vertical residual velocity (Fig. 7k and blue line in Fig. 8). That is, rising motion balances in part the shortwave heating in the tropical upper stratosphere. The balance among these terms as a function of time after branching is shown in Fig. 8, which indicates that over the first 15 days after branching, the change in diabatic heating (J/C_p) mostly leads to warmer temperatures, but after approximately

day 15 the dominant balance is between diabatic heating and upwelling [$(N^2 H/R)\bar{w}^*$] with little additional temperature increase ($\partial T/\partial t \rightarrow 0$). Anomalous subsidence occurs in midlatitudes (Fig. 6a), and this subsidence drives local warming that allows for the warm anomaly to spread out from the subtropics and into midlatitudes (Fig. 7b). This warming of the midlatitudes leads to longwave cooling that becomes more apparent at later lags (e.g., right column of Fig. 7). The key point from the TEM heat budget is that the shortwave heating increase for enhanced UV directly leads to an accelerated BDC, with enhanced upwelling in the tropical upper stratosphere (Fig. 6 and blue line in Fig. 8) and downwelling in midlatitudes. However this direct response is dwarfed, after a few weeks, by changes in the BDC that are not driven by the shortwave heating directly.

What leads to these changes in the BDC a few weeks after branching? To understand the lagged response, we turn our attention to the TEM zonal momentum balance displayed in Eq. (2). In Fig. 9 we show the right-hand terms from the zonal momentum equation in the first two days, to demonstrate how they relate to the zonal wind acceleration immediately after branching. The first important result is that the budget largely closes. Figure 9g shows the wind tendency over the first two days, while Fig. 9h shows the sum of all forcing terms on the right-hand term of Eq. (2). While there are some dissimilarities between them, especially near the equator where $u'w'$ may contribute to EPF, the subpolar maxima are accounted for by the momentum budget. Hence the momentum budget can be used to diagnose which specific process leads to strengthened winds.

We next consider each term in the TEM zonal momentum balance, in order to understand (i) how the UV forcing leads to wind anomalies in subpolar latitudes where there is little sunlight and (ii) how a weakened BDC develops at later lags (Figs. 5 and 6). The enhanced vertical residual velocity in the tropics and subsidence in midlatitudes are connected by enhanced poleward flow in the subtropical upper stratosphere in order to maintain mass continuity (Fig. 6), as would be expected given the theoretical arguments of Semeniuk and Shepherd (2001). This poleward flow leads to an acceleration that peaks near 30°N in the first two days after branching (Fig. 9b) via Coriolis torques. This acceleration does not reach subpolar latitudes, and specifically is near-zero near 60°N . The $f\bar{v}^*$ term dominates the first two days after branching, although eddies, as diagnosed by Fig. 9a, have already begun to respond and lead to an acceleration poleward of 60°N (and deceleration farther equatorward), such that there is a net acceleration throughout the NH upper stratosphere already in the first two days. Over the first two days, the remaining three terms are not important (Figs. 9d–f).

During days 3–9 after branching (Fig. 10), a dipole in Eliassen–Palm flux divergence (EPFD) develops: westerly momentum is fluxed out of the subtropical upper stratosphere (i.e., there is EPF convergence in the region where $f\bar{v}^*$ drives an acceleration) and is fluxed toward subpolar latitudes (i.e., EPF divergence anomalies) where the direct, radiative response to solar variability is weak (Fig. 10a). The EPF divergence in subpolar latitudes is in turn balanced by an anomalous equatorward residual velocity at subpolar latitudes from 20 to 5 hPa, corresponding to the decelerated BDC in the polar

FIG. 7. Difference between enhanced UV and reduced UV branch runs for terms in the TEM thermodynamic equation [Eq. (4)] in (a)–(c) days 1 and 2 after branching, (d)–(f) days 3–9 after branching, (g)–(i) days 10–19 after branching, and (j)–(l) days 20–29 after branching. Dots indicate where a null hypothesis of no difference can be rejected at the 5% confidence level. Note that the contour interval differs for the first row. Shown are (left) temperature tendency, (center) dynamical heating and meridional advection, and (right) diabatic heating.

FIG. 8. TEM thermodynamic budget terms [Eq. (4)] and their sum in the tropical stratosphere from 7°S to 7°N and pressure weighted 1–7 hPa: sensible temperature change ($\partial\bar{T}/\partial t$; red); temperature tendency due to radiation (J/C_p ; green); vertical residual velocity variations [$(N^2 H/R)\bar{w}^*$; blue]; and the sum of the three most important terms in Eq. (4): $\partial\bar{T}/\partial t + (N^2 H/R)\bar{w}^* - J/C_p$ (black).

region (Fig. 10b). The net effect of the change in $f\bar{v}^*$ and EPFD is an acceleration of winds near the polar vortex (Fig. 10c). Enhanced EPF convergence in the subtropics is associated with an enhanced BDC, while reduced EPF convergence in subpolar latitudes is associated with a weakened BDC (Figs. 6b,c).

On days 10–19 (Fig. 11), the EPF divergence at high latitudes and the EPF convergence in the subtropics continue to strengthen, leading to a flux of momentum out of the subtropics (where $f\bar{v}^*$ drives an acceleration) and toward the pole. Hence eddies play a crucial role in accelerating the vortex at latitudes where there is little sunlight. The EPFD also is associated with changes in $f\bar{v}^*$ in the subpolar midstratosphere, and specifically a weakened BDC. The accelerated BDC that originally extended to the lower stratosphere and to the pole is now confined to the tropical and subtropical stratosphere on days 10–19 (Figs. 11b and 6c) and to the tropical uppermost stratosphere on days 20–29 (Figs. 12b and 6d).

Other terms in the TEM momentum budget also contribute on days 10–19 and 20–29. The vertical advection (Fig. 11d) acts to weakly decelerate the vortex in polar latitudes in the mid-stratosphere by upward advection of weaker westerly winds from lower altitudes, but it accelerates the vortex between 40° and 60°N due to anomalous downwelling associated with a weakened BDC at polar latitudes. The meridional advection term accelerates the stratospheric vortex poleward of 60°N by downgradient advection from the vortex core at ~60°N where winds are strongest, but decelerates the vortex farther equatorward. Both of the advection terms intensify significantly as the negative subpolar streamfunction anomaly develops, and for days 20 to 29 the acceleration due to meridional advection is comparable in magnitude to the EPFD itself, and exceeds $2 \text{ m s}^{-1} \text{ day}^{-1}$ between 60° and 80°N extending vertically throughout the entire stratosphere (Fig. 12e). Even the gravity wave drag takes part in forcing an acceleration of extratropical winds, although its

contribution is weaker compared to the other terms and is confined to the upper-most stratosphere. A partial cancellation of changes in resolved waves (i.e., EPFD) by changes in gravity wave drag is consistent with previous work (Cohen et al. 2013; Sigmond and Shepherd 2014; Scheffler and Pulido 2015; Garfinkel and Oman 2018).

Overall, the subtropical poleward flow associated with the acceleration of the BDC in the tropical upper stratosphere leads to accelerated zonal winds. This acceleration of subtropical zonal winds is partially balanced by EPF convergence in the subtropics and accompanying EPF divergence over the poles, which fluxes westerly momentum poleward and broadens the region experiencing intensified westerlies. This EPF divergence over the poles necessitates weaker subsidence over the poles than in the climatology (i.e., anomalous rising motion) that is associated with adiabatic cooling over the poles. The zonal wind acceleration then extends downward to the lower stratosphere.

This change in UV radiation also affects stratospheric sudden warming frequency. The control run simulates 3.8 sudden warmings per decade using the Charlton et al. (2007) definition, and over the 29 years used here for initializing the altered UV ensembles eight sudden warmings occurred between 1 January (the day on which we branch) and 15 March. The enhanced UV branch ensemble also includes eight runs with a sudden warming between 1 January and 15 March, while the reduced UV ensemble includes 11 branch runs with a sudden warming between 1 January and 15 March. While this slight increase in sudden warming frequency for reduced UV is consistent with the weakened vortex, it is not statistically significant. However if we focus only on sudden warmings between 1 February and 15 March after the polar response has mostly developed, the difference among the three ensembles is more pronounced: three enhanced UV and seven reduced UV ensemble members simulate a sudden warming out of 29 total members, while four sudden warmings occurred in the control run over the corresponding period.

How long does it take for the UV perturbation to affect tropospheric conditions? In Fig. 13 we focus on the downward propagation of the stratospheric zonal wind anomaly. The zonal wind anomaly at 100 hPa first appears roughly two weeks after branching, but at 300 and 700 hPa strong anomalies only appear after approximately a month. The dipole anomaly across 40°N at the 700- and 300-hPa levels indicates a northward shift of the tropospheric jets (Haigh et al. 2005; Simpson et al. 2009), supporting the notion of a stratospheric pathway for solar variability to influence surface climate. The time scale over which a lower stratospheric response develops is longer for the ensemble of branch runs initialized on 1 November (Fig. S8), consistent with the delayed lower stratospheric response (Figs. S4 and S5), and a tropospheric response is evident only 50 days after branching. Implications for observed solar variability are discussed next in section 4.

4. Discussion of possible implications for the observed response to solar variability

The UV forcing imposed here is approximately a factor of 10 larger than any observed change over the satellite record, and

FIG. 9. Difference between increased UV and decreased UV branch runs for terms in the TEM momentum equation on days 1 and 2 after branching. Dots indicate where a null hypothesis of no difference can be rejected at the 5% confidence level. The contour interval is $0.4 \text{ m s}^{-1} \text{ day}^{-1}$. Shown are (a) $(\nabla \cdot \mathbf{F})/(\rho_0 a \cos \phi)$, with EP flux arrows shown in gray if either component is statistically significant; (b) fv^* ; (c) the sum of (a) and (b); (d) $-w^*(\partial \bar{u}/\partial z)$; (e) $-[(\bar{u} \cos \phi)_\phi / a \cos \phi] \bar{v}^*$; (f) gravity wave drag; (g) $\partial \bar{u}/\partial t$; and (h) the sum of the right-hand side of Eq. (2).

is also imposed stepwise while solar UV radiation changes gradually. We view these unrealistic aspects of the forcing as a positive attribute however, as an exaggerated, stepwise forcing helps to clarify the development of the important dynamical mechanisms immediately after the forcing is switched on (as is now done by dozens of modeling centers in the abrupt $4\times\text{CO}_2$ experiment included in CMIP6). That being said, this lack of realism in the forcing makes it difficult to directly compare our results to the observed response to solar variability over the satellite era. These important caveats notwithstanding, our results may have implications for the impact of observed solar variability, as we now discuss.

We find no evidence for a secondary lower stratospheric maximum in temperature even though the BDC is weakened in the tropical lower stratosphere after day 20 (Fig. 6). Namely, the temperature anomaly increases monotonically upward in the tropics over the entirety of the branch runs. Hence a

weakened BDC does not necessarily require a secondary lower stratospheric maximum in temperature, and some other mechanism (or aliasing from some other source of variability) is likely associated with the observed secondary maximum. For example, ozone is fixed in our experiments, and it is conceivable that the lack of ozone feedbacks in our model may be important for the lack of a secondary lower stratospheric maximum.

The top-down mechanism discussed in section 3 involving the extratropical stratosphere likely cannot explain any apparent connection between the 27-day solar rotation period and surface climate, as the zonal wind response in the troposphere in Fig. 13 is delayed by a month. However, this mechanism might explain a connection of surface climate with low-frequency variations in the solar forcing on paleoclimate time scales. Note that the internal stratospheric time scales associated with SSWs in this model as compared to reanalysis

FIG. 10. As in Fig. 9, but for days 3–9.

were examined by White et al. (2020) and found to be realistic (their Fig. 2a). The time scales for the tropospheric response and extratropical stratospheric response for the 1 November ensemble is even longer than for the 1 January ensemble, likely because the upward wave flux is weaker in November than in January, and hence the eddy feedbacks develop more slowly. The inability of the 27-day solar rotation period to initiate the entirety of the top-down mechanism has also been suggested by two previous studies. Garfinkel et al. (2015), using reanalysis data, found extratropical impacts from the 27-day solar rotation in the upper and middle stratosphere, but not lower down, and Gruzdev et al. (2009) similarly found a weak dynamical response in a chemistry climate model. Nevertheless, it is worth revisiting this question with future reanalysis data and comprehensive model output.

The model used here does not represent all potential mechanisms that have been proposed to be important for surface impacts from the 27-day solar rotation period. Specifically, the model does not spontaneously develop a Madden–Julian oscillation (not shown) and hence cannot be used to consider the

question of whether the 27-day solar rotation period leads to an extratropical response by modifying the Madden–Julian oscillation (Hood 2018). In addition, Haigh et al. (2005), Simpson et al. (2009), and Simpson et al. (2010) argue that the secondary lower stratospheric maximum leads to an immediate deceleration on the equatorward flank of the tropospheric jet that resembles its equilibrium response by days 20–29 after the anomalous heating is imposed (i.e., they do not simulate the stronger UV flux explicitly). While we find a generally similar tropospheric jet shift, it only occurs after at least a month in our simulations and is associated with changes in the subpolar lower stratosphere. It is conceivable that a model that simulates a secondary lower stratospheric maximum could simulate a tropospheric response faster than we find here.

Finally, we find that the stratospheric response is present for either reduced or enhanced UV, although some deviations from a linear response are present (Figs. 2 and 3). While our results are likely to be relevant for smaller perturbations as well, the response to an observed UV perturbation will suffer

days 10 to 19, enhanced–reduced UV

FIG. 11. As in Fig. 9, but for days 10–19.

from a low signal-to-noise ratio as discussed by Sukhodolov et al. (2017).

5. Summary

Variability in incoming solar radiation impacts the stratosphere in both the tropics and the winter pole where there is no sunlight, and the response is summarized schematically in Fig. 14. Tropical stratospheric warming follows immediately after incoming UV is increased, consistent with previous studies (Fig. 14a; Swartz et al. 2012; Bednarz et al. 2019, among others). The tropical upper stratospheric warming leads to enhanced subtropical zonal winds via thermal wind balance, and these subtropical zonal wind anomalies shift poleward in time due to changes in wave propagation (i.e., changes in EPFD; Fig. 14a), ultimately leading to a stronger polar vortex that slowly develops over the first month after incoming UV is increased. Only after approximately a month is a robust effect evident in the troposphere: the tropospheric jet shifts poleward in response to enhanced UV (Fig. 14b).

Increased shortwave heating due to enhanced incoming solar UV is associated with enhanced rising motion in the tropical upper stratosphere. Anomalous subsidence then occurs in the midlatitude upper stratosphere, leading to local dynamical warming. Poleward motion that occurs in this cell drives an acceleration of westerlies in the subtropical winter stratosphere via Coriolis torques. These changes in the subtropics are accompanied by increased Eliassen–Palm flux convergence in the subtropics and divergence in the subpolar latitudes, which leads to a poleward extension of the enhanced westerlies and a weakened BDC in the subpolar middle and lower stratosphere. Wave fluxes also lead to the downward propagation of the signal to the lower stratosphere and then into the troposphere, and the weakened BDC expands to all regions of the winter stratosphere except near the tropical stratopause.

The stratospheric response to the altered UV includes a modulation of the BDC, and the changes in the BDC are nonmonotonic in most of the stratosphere. An accelerated residual circulation in response to enhanced UV and shortwave heating occurs throughout the stratosphere in the first

days 20 to 29, enhanced–reduced UV

FIG. 12. As in Fig. 9, but for days 20–29.

two days (Fig. 14a), but only in the tropical upper stratosphere is this acceleration maintained throughout the duration of the integrations. Elsewhere, the initial BDC acceleration is replaced by a weakened residual circulation in response to enhanced UV. This later weakening of the BDC is dynamically driven, and specifically is associated with reduced wave convergence in subpolar latitudes in response to enhanced UV (Fig. 14b). According to Tung and Kinnersley (2001) and Semeniuk and Shepherd (2001), an enhanced annular momentum transfer into the winter subtropical upper stratosphere (anomalous EPFD convergence as is indeed evident in Figs. 9 to 12) would result in strengthened tropical upwelling that extends into the middle stratosphere and cross-equatorial flow in the upper stratosphere (as is indeed evident in Fig. 6), although downwelling occurs in the lower stratosphere after day 20.

The switch-on experiments allow us to estimate the time scale of the stratospheric response in every region. The tropical temperature response, for example, occurs mostly within the first 10 days and then stabilizes, while at high latitudes the zonal

wind anomaly continues to increase over the first month. In the 1 January ensemble, changes in wind at 100 hPa only occur ~ 15 days after branching, and changes in the lower troposphere begin to appear only after a month.

An unanswered question in this study is this: How does the EPFD dipole initially develop in the upper stratosphere that shifts the initial subtropical acceleration poleward and ultimately leads to a polar stratospheric response? The quasi-geostrophic index of refraction [computed as in Matsuno (1970)] for stationary waves initially decreases in the subtropical upper stratosphere in response to enhanced UV, which, if anything, should indicate that waves are ducted away from the subtropics (Fig. 15). Even though the meridional gradient of potential vorticity increases in this region in response to enhanced UV, which by itself would help wave propagation (not shown), the increase in the zonal wind (which appears in the denominator of the expression for the index of refraction) is even larger and leads to a reduction in index of refraction even as wave propagation into this region is apparently enhanced. Nevertheless, the tendency for waves to flux

FIG. 13. Difference between increased UV and decreased UV branch runs in the zonal wind as a function of time after branching, at the (a) 100-, (b) 300-, and (c) 700-hPa level. Contour interval is 0.5 m s^{-1} . The climatological tropospheric jet maximum is indicated with gray x marks in (b) and (c).

momentum out of a region where an external perturbation is applied is consistent with the results of [Watson and Gray \(2015\)](#). That linear theory seems to not fully describe the subpolar response to solar variations may provide support for the nonlinear mechanisms described in [Lu et al. \(2017a,b\)](#).

There are some indications that the extratropical stratospheric response to the solar cycle depends on the phase of the quasi-biennial oscillation ([Labitzke 1987](#); [Labitzke et al. 2006](#); [Garfinkel et al. 2015](#); [Rao et al. 2019](#)). A thorough exploration of how the quasi-biennial oscillation may affect the results presented here is left for future work.

Schematic of the response to enhanced UV

FIG. 14. Schematic of the influence of altered UV radiation on the winter atmosphere after (a) 2 days and (b) 20 days. Gray lines represent changes in the TEM residual streamfunction. The thick red arrow indicates that warm anomalies are spreading poleward in time, while the thick green arrow indicates that westerly anomalies are propagating downward in time.

Matsuno 1970 Refractive index, enhanced-reduced UV

FIG. 15. As in Fig. 4, but for anomalies of the index of refraction from the control run. The contour interval is 1.5.

Overall, these results support the existence of the stratospheric pathway, and help resolve the apparent contradiction between enhanced shortwave heating with a weakened BDC during solar maximum. However, any tropospheric impact is only apparent approximately a month after the solar forcing is altered, and hence the specific top-down mechanism focused on in this work may be too slow to account for an extratropical tropospheric response to solar variability on the 27-day solar rotation time scale.

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APPENDIX

Implementation of Modified UV in MiMA

The shortwave Kurucz source function fed into MiMA is made up of 29 intensity bands, corresponding to wavelengths of

3000 and down to 200 nm. Bands 27 and 28, corresponding to wavelengths 260–340 and 260–200 nm, are multiplied by a factor of 4/3 for increased UV, and by a factor of 2/3 for decreased UV.

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