The Influence of Aerosol Absorption on the Extratropical Circulation

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ABSTRACT

This study examines how aerosol absorption affects the extratropical circulation by analyzing the response to a globally uniform increase in black carbon (BC) simulated with an atmospheric general circulation model forced by prescribed sea surface temperatures. The model includes aerosol direct and semidirect effects, but not indirect or cloud-absorption effects. BC-induced heating in the free troposphere stabilizes the midlatitude atmospheric column, which results in less energetic baroclinic eddies and thus reduced meridional energy transport at midlatitudes. Upper-tropospheric BC also decreases the meridional temperature gradient on the equatorward flank of the tropospheric jet and yields a weakening and poleward shift of the jet, while boundary layer BC has no significant influence on the large-scale circulation since most of the heating is diffused by turbulence in the boundary layer. The effectiveness of BC in altering circulation generally increases with height. Dry baroclinic eddy theories can explain most of the extratropical response to free-tropospheric BC. Specifically, the decrease in vertical eddy heat flux related to a more stable atmosphere is the main mechanism for reestablishing atmospheric energy balance in the presence of BC-induced heating. Similar temperature responses are found in a dry idealized model, which further confirms the dominant role of baroclinic eddies in driving the extratropical circulation changes. The strong atmospheric-only response to BC suggests that absorbing aerosols are capable of altering synoptic-scale weather patterns. Its height dependence highlights the importance of better constraining model-simulated aerosol vertical distributions with satellite and field measurements.

1. Introduction

The large-scale atmospheric circulation response to climate forcings has been studied extensively. The greenhouse gas (GHG)-induced warming is thought to cause a poleward shift of the subtropical jets and storm tracks and an expansion of the tropics (e.g., Hall et al. 1994; Yin 2005; Lorenz and DeWeaver 2007; Lu et al. 2008; Chen et al. 2008). Scattering aerosols (e.g., sulfate) can partly offset the climate impacts of GHGs by reflecting solar radiation. Previous studies using coupled general circulation models (GCMs) have shown the impacts of aerosols on both tropical and extratropical circulation. As a result of the interhemispheric asymmetry in the aerosol forcing, the Hadley circulation weakens (strengthens) in the boreal summer (winter) and the intertropical convergence zone shifts southward

(Ming and Ramaswamy 2011). Aerosol-induced cooling results in an equatorward shift of the jet stream, opposite to the GHG-induced change (Fischer-Bruns et al. 2009; Ming and Ramaswamy 2009). Allen and Ajoku (2016) showed that future projected aerosol decrease results in tropical widening and a poleward shift of the jet, and the effect of aerosol decrease may be larger than GHG increase in GCMs, including aerosol indirect effect. Ming et al. (2011) suggested that aerosols also cause zonal-asymmetric circulation change at midlatitudes by altering stationary Rossby waves, which results in a strong cooling and a decrease of transient eddy kinetic energy (EKE) over the North Pacific. On regional scales, aerosols have been shown to modify radiative heating rates at the surface and in the boundary layer in urban areas and thus affect the temperature profile (Ackerman 1977; Jacobson 1998). Jacobson and Kaufman (2006) suggested that aerosols tend to reduce near-surface wind speed and precipitation locally by stabilizing the

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atmosphere. Recent studies using atmospheric GCMs (AGCMs) have shown that anthropogenic aerosols also modulate midlatitude cyclones by changing the vertical profile of diabatic heating rates in the atmosphere (e.g., Wang et al. 2014a,b; Lu and Deng 2016).

Absorbing aerosols (e.g., black and brown carbon) have different radiative properties from scattering aerosols and contribute to global warming along with GHGs. Black carbon (BC) from fossil fuel emissions has been shown to increase the stability of near-surface air, slow near-surface winds, and reduce global mean precipitation (e.g., Jacobson 2002, 2004; Ming et al. 2010). Some studies have suggested that controlling BC emissions may be more effective at slowing global warming than controlling GHG emissions (Jacobson 2002, 2010). Recent work has focused on the effects of absorbing aerosols on regional climate (e.g., Bollasina et al. 2008; Randles and Ramaswamy 2008). However, the impacts of absorbing aerosols on atmospheric circulation have received little attention. Allen and Sherwood (2011) showed that the circulation responses to natural (mostly scattering) and anthropogenic (scattering and absorbing) aerosols are very different and inferred that absorbing aerosols strongly affect atmospheric circulation in an opposite way to scattering aerosols. While both absorbing aerosols and GHGs act to warm the climate, their effects on large-scale circulation are not necessarily similar. Ming et al. (2010) showed that the global mean precipitation increase resulting from the warming caused by absorbing aerosols does not scale with surface temperature change since the strong atmospheric absorption suppresses precipitation. Using the Community Atmosphere Model coupled to a slab ocean, Allen et al. (2012a) showed that BC and tropospheric ozone play a more important role than GHGs in driving tropical expansion in the Northern Hemisphere in recent years because of the associated atmospheric heating at mid-latitudes and the resulting poleward shift of the maximum meridional temperature gradient. Despite these early attempts, the influence of absorbing aerosols on large-scale circulation has not been studied systematically.

In general, global emissions of BC have increased in recent decades while sulfate emissions have declined (Lamarque et al. 2010), and this trend is projected to continue over the next decade under several representative concentration pathways (Lamarque et al. 2011; Fiore et al. 2012). This adds urgency to understanding the circulation response to absorbing aerosols for attributing the observed trend and variability in atmospheric circulation and predicting future changes. Satellite and in situ observations show that a large amount of BC is present both in the tropics and at midlatitudes (e.g., Koch et al. 2009; Schwarz et al. 2013). One would expect that the

circulation response to the same forcing varies with latitude because of different dynamical regimes. In the tropics where the Coriolis parameter is small, the time mean flow is the largest contributor to the poleward energy transport. The vertical temperature structure (or the static stability) is set approximately by the moist adiabat. In the extratropics, baroclinic eddies play the dominant role in transporting heat and moisture poleward and shaping the large-scale circulation and weather pattern. The static stability is largely controlled by dry baroclinic eddy dynamics (Held 1982; Zurita-Gotor and Lindzen 2007; Schneider and O’Gorman 2008), while moisture has an important but secondary role (Frierson 2008). In light of the very different tropical and extratropical regimes, in this study we choose to focus on the impacts of absorbing aerosols on the extratropical circulation and associated physical mechanisms.

The overall picture of aerosol–climate interactions is complicated and uncertain since it involves a variety of physical processes, such as aerosol emission and transport, aerosol–radiation interactions, aerosol–cloud interactions, and air–sea coupling. To simplify the problem, we use an AGCM to study tropospheric-only response to idealized absorbing aerosol forcings. Our analysis includes only aerosol direct and semidirect effects. We examine the effects of absorbing aerosols at different altitudes since previous studies have shown that BC–climate interaction is highly dependent on the vertical profile (Hansen 2005; Ming et al. 2010; Ban-Weiss et al. 2012; Persad et al. 2012).

2. Method

We use the Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model version 2 (AM2), the atmospheric component of GFDL Climate Model version 2 (CM2), to investigate the atmospheric-only response to absorbing aerosols. The configuration and performance of this model have been documented in Anderson et al. (2004); here we describe briefly the features most relevant to this study. AM2 uses a finite volume dynamical core with a horizontal resolution of about $2^\circ \times 2.5^\circ$. The model has 24 hybrid vertical levels from the surface to 3 hPa, with 9 levels in the planetary boundary layer (the lowest 1.5 km), 10 levels in the free troposphere, and 5 levels in the stratosphere. The shortwave and longwave radiation algorithms follow Freidenreich and Ramaswamy (1999) and Schwarzkopf and Ramaswamy (1999), respectively, with modifications as described in Anderson et al. (2004). The model uses the relaxed Arakawa–Schubert (RAS) convective parameterization, which represents moist convection as multiple entraining plumes that produce precipitation (Moorthi and Suarez 1992). Stratiform clouds are

forecast following Tiedtke (1993), with modifications as described in Anderson et al. (2004). Cloud microphysics is parameterized based on Rotstajn (1997) and Rotstajn et al. (2000). Convective planetary boundary layers are parameterized using a K-profile scheme based on Lock et al. (2000). For stable layers, conventional stability functions dependent on the Richardson number are used. Tropospheric aerosols and ozone are simulated offline using a chemical transport model driven by GCM-simulated meteorological fields (Horowitz 2006). All aerosols are treated as externally mixed, and lognormal size distribution is assumed with the geometric mean radius and standard deviation from Haywood and Ramaswamy (1998). The model includes only the direct effects of aerosols with optical properties described by Haywood and Ramaswamy (1998) and Haywood et al. (1999). The model has been used to study the responses of general circulation and hydrological cycle to GHGs and aerosols (e.g., Ming et al. 2010; Persad et al. 2012).

We perturb the control case (with GHG and aerosol concentrations in 1990) with an increase of $2.4 \times 10^{-6} \text{ kg m}^{-2}$ in BC burden within a specific model layer over the entire globe. This burden is chosen so that the resulting radiative forcing is comparable to that of the present-day BC [approximately 0.53 W m^{-2} in AM2 (Ming et al. 2010)]. The BC burden is prescribed and not interactive. The globally uniform increase in BC is not representative of present-day BC or any future scenario. Given large uncertainties in the realistic spatial distribution of aerosols, we choose to use these uniform absorbing aerosol experiments to investigate the underlying mechanisms through which the climate impacts are manifested. We examine the model-simulated response to increase of BC at three model sigma layers in the free troposphere or the boundary layer ($\sigma = 0.38, 0.60$, and 0.90). To investigate to what extent the response is local, we also perform experiments with latitudinally restricted increase of BC in the tropics (30°S – 30°N), midlatitudes (30° – 60°N and 30° – 60°S), and high latitudes (60° – 90°N and 60° – 90°S) at $\sigma = 0.38$. The simulations are forced with monthly climatological sea surface temperatures (SSTs) and sea ice from the NOAA Optimum Interpolation Sea Surface Temperature dataset (Reynolds et al. 2002). Each simulation is run for 17 years, and the results in this paper are averaged over the last 16 years.

We also conduct a set of idealized model experiments to complement the comprehensive model results. The idealized model is based on a spectral dynamical core. It uses a sigma coordinate, with the vertical differencing following Simmons and Burridge (1981). The model is dry (i.e., has no clouds or water vapor) and has neither topography nor seasonal cycle. It is forced with highly idealized physics as described in Held and Suarez

(1994). Radiative heating and cooling are represented by Newton relaxation of temperature to a specified zonally symmetric radiative-equilibrium state. Momentum is damped by Rayleigh friction near the surface, the rate of which decreases linearly from 1 day^{-1} at the surface to 0 day^{-1} at $\sigma = 0.7$. The model does not have parameterized convection. This dry idealized GCM has been used to study the response of tropospheric circulation to idealized forcings such as stratospheric warming, surface friction, and zonal torque (Chen et al. 2007; Lorenz and DeWeaver 2007; Chen and Zurita-Gotor 2008). We run the model with a horizontal resolution of spectral T85 ($\sim 1.4^\circ$) and 30 evenly spaced vertical levels, and we perturb the control run by adding a global uniform heating rate of $3 \times 10^{-5} \text{ K s}^{-1}$ at specific levels. The heating rate is chosen to yield an anomalous column-integrated heating rate similar to that induced by BC in the comprehensive model. This is done for two different sigma layers in the free troposphere ($\sigma = 0.38$ and 0.58). The model is integrated for 1000 days for each experiment, and the last 500 days are used for analysis.

3. Results

a. Temperature and zonal wind

Figure 1 shows the responses of zonal mean temperature and zonal wind to BC at different levels. In general BC heats the troposphere by absorbing solar radiation, and the maximum warming occurs at heated altitudes. The temperature increase caused by BC at higher altitudes is much stronger. Upper- ($\sigma = 0.38$) and mid-tropospheric ($\sigma = 0.60$) BC yields a maximum warming of about 6 and about 3 K, respectively, while the temperature change resulting from boundary layer ($\sigma = 0.90$) BC is less than 1 K. The magnitude of warming decays away throughout the troposphere, which stabilizes (destabilizes) the atmosphere below (above) the heating layer. The tropospheric warming penetrates to lower altitudes more in the midlatitudes than in the low and high latitudes, indicating that the atmosphere responds differently in the three distinct dynamical regimes. More specifically, the tropical air temperature is under strong control of the surface temperature through moist convection; the latter is fixed in our simulations. The stable polar atmosphere is not conducive to vertical mixing. A detailed discussion of the midlatitudes will be given later. Midtropospheric and boundary layer BC is more effective at exciting surface polar amplification. Free-tropospheric BC also results in cooling in the polar stratosphere and warming near the tropopause in the tropics. These nonlocal responses may result from the stratospheric residual circulation change, which is out of the scope of this paper.

FIG. 1. Changes in (left) zonal mean temperature and (right) zonal wind resulting from BC at $\sigma =$ (a),(d) 0.38, (b),(e) 0.60, and (c),(f) 0.90. The black contours denote the climatological mean from the control run. The hatching represents significance at the 0.05 level.

Upper-tropospheric BC has a strong effect on both the subtropical jet and the eddy-driven jet, which merge together in the climatological control run (Fig. 1d). There is an appreciable weakening of the zonal wind on the equatorward flank of the subtropical jet ($\sim 20^\circ$ latitude) which is accompanied by a strengthening on the poleward flank of the eddy-driven jet ($\sim 60^\circ$ latitude). If one defines the jet position as the latitude of maximum vertical averaged zonal mean zonal wind, this wind pattern change amounts to a poleward jet displacement of about 3° latitude in both hemispheres. The jet response is a result of changes in both the vertical wind shear and surface wind. The vertical shear decreases (increases) on the equatorward (poleward) flank of the jet, consistent with the anomalous meridional temperature gradient (Fig. 1a). The poleward shift of surface westerlies is related to the change in eddy momentum flux, which is shown in Fig. 2. In both hemispheres, the decrease (increase) in eddy

momentum flux on the equatorward (poleward) flank of the jet gives rise to a divergence of eddy momentum flux at midlatitudes, which slows down surface westerlies. In contrast, the convergence of eddy momentum flux poleward of about 60° latitude acts to accelerate surface westerlies. The negligible jet displacement in the case of midtropospheric BC is likely due to the competing effects of the surface warming amplification at high latitudes and the upper-tropospheric warming amplification in the tropics. This is consistent with previous studies focusing on the jet shift under global warming, which have shown that increased upper-tropospheric meridional temperature gradient tends to shift the jet poleward while decreased lower-tropospheric temperature gradient does the opposite (e.g., Barnes and Screen 2015; Butler et al. 2010). The impact of boundary layer BC on zonal wind is similar to that of midtropospheric BC, albeit with an even smaller magnitude, especially in the Northern Hemisphere.

FIG. 2. Changes in zonal mean eddy momentum flux resulting from BC at $\sigma = 0.38$. The black contours denote the climatological mean from the control run. The hatching represents significance at the 0.05 level.

Figures 3a–c show the responses of zonal mean temperature in the latitudinally restricted perturbation experiments in which BC is increased at $\sigma = 0.38$ in the tropics, midlatitudes, and high latitudes, respectively. Increased BC at individual latitude bands yields maximum warming at the heated latitudes. To first order the temperature response is local; the warming is mostly confined in the forced latitudinal bands. It is notable that tropical BC also causes some dynamically induced warming at midlatitudes. In contrast to the clear downward mixing in the midlatitude case, the high-latitude warming is almost entirely confined locally, a manifestation of the stable atmospheric condition in the polar regions. A comparison with Fig. 1a suggests that the temperature response to BC at different latitudes is mostly linearly additive.

Figures 3d–f depict the responses of zonal wind to the latitudinal restricted forcings, with corresponding changes in eddy momentum flux in Fig. 4. The poleward jet displacement seen in the case of globally uniform BC can be attributed mostly to midlatitude BC (Fig. 3e). The resulting divergence (convergence) of eddy momentum flux decelerates (accelerates) surface westerlies equatorward (poleward) of 60° latitude. Upper-tropospheric wind anomalies are consistent with changes in the meridional temperature gradient and the associated vertical wind shear. The poleward jet shift caused by midlatitude BC is more prominent than that in the globally uniform case in the Southern Hemisphere. This is mainly because high-latitude BC has an opposite effect, reducing zonal wind on the poleward flank of the jet and yielding an equatorward jet displacement (Fig. 3f). The opposite effects of mid- and high-latitude heating on the jet have also been found in GCM experiments with idealized thermal forcings at different latitudes (Allen et al. 2012b). Tropical

BC results in an anomalous poleward eddy momentum flux at the jet core in both hemispheres, which helps force the weakening (strengthening) of the surface wind near 30° (60°) latitude. In the upper troposphere the eddy-driven jet becomes stronger, consistent with the increased meridional temperature gradient as a result of tropical warming. In the Southern Hemisphere tropical BC also results in a slight poleward jet displacement, but midlatitude BC is much more effective at shifting the jet.

b. Mean circulation and eddy activity

Figure 5 shows the response of the meridional overturning streamfunction to BC at different altitudes. Upper-tropospheric BC results in a weakening and expansion of the Hadley cell in both hemispheres. The weakening occurs in the summer hemisphere of the solstice seasons and in both hemispheres of the equinox seasons (not shown). This is related to the anomalous eddy momentum flux convergence in the upper troposphere at about 20° latitude (Fig. 2), consistent with the linear theories of Hadley circulation strength (e.g., Walker and Schneider 2006). The Hadley cell expansion may be due to the increase in subtropical static stability (Fig. 1a), as suggested by the existing scaling theories of Hadley circulation extent (e.g., Walker and Schneider 2006; Lu et al. 2007). In the extratropics upper-tropospheric BC results in a weakening of the Ferrel cell. This is consistent with the change in eddy momentum flux (Fig. 2), as the anomalous divergence of the eddy momentum flux in the upper troposphere at midlatitudes is balanced by the Coriolis torque acting on the anomalous poleward flow. Midtropospheric and boundary layer BC yields a similar weakening of the Hadley and Ferrel cells, but the magnitude is much smaller.

Atmospheric circulation plays an important role in transporting energy from equatorial regions to higher latitudes. This poleward energy flux occurs mainly through the mean meridional circulation, stationary eddies, and transient eddies. Figure 6 shows the change in total northward energy flux resulting from BC at different altitudes and the contribution from each component. In the tropics the weakening of the Hadley cell resulting from free-tropospheric BC results in a decrease in the poleward energy transport by mean circulation. In the extratropics free-tropospheric BC causes a decrease in energy transport by transient eddies. The weakening of the energy transport occurs everywhere below the heating layer (Figs. 7a,b). Overall, the poleward energy flux by transient eddies decreases by about 14% (5%) at midlatitudes because of upper-tropospheric (midtropospheric) BC. In the Northern Hemisphere part of the decrease is balanced by an anomalous northward energy flux associated with the weaker Ferrel cell

FIG. 3. Changes in (left) zonal mean temperature and (right) zonal wind resulting from BC in the (a),(d) tropics (30°S – 30°N), (b),(e) midlatitudes (30° – 60°N and 30° – 60°S), and (c),(f) high latitudes (60° – 90°N and 60° – 90°S) at $\sigma = 0.38$. The black contours denote the climatological mean from the control run. The hatching represents significance at the 0.05 level.

(Figs. 5a,b), but in general transient eddies dominate the weakening of poleward energy transport in both hemispheres. The stationary eddies have seasonal variations with opposite signs in summer and winter (not shown) and thus do not contribute to the change in annual mean meridional energy flux. The change in meridional energy flux resulting from boundary layer BC is not statistically significant in most places (Figs. 6c and 7c).

Since the poleward eddy transport of energy at midlatitudes can be thought of as turbulent diffusion (Held 1999), the anomalous energy flux is related to the changes in the meridional gradient and the eddy strength. Further calculations suggest that the meridional moist static energy gradient at midlatitudes does not change significantly; thus, the decrease in the energy flux is caused by weaker eddies. To understand the change in eddy activities, we examine the velocity scale V and the

length scale L of the baroclinic eddies. Figures 7d–f show the change in EKE resulting from BC at different altitudes. Free-tropospheric BC results in a reduction in EKE, which peaks in the upper troposphere where the climatological EKE is the strongest. The average velocity of the eddies (square root of the mean EKE) decreases by about 13% (3%) because of upper-tropospheric (mid-tropospheric) BC. The change in EKE resulting from boundary layer BC is, again, not statistically significant. The decrease in eddy velocity is largely a result of the enhanced static stability, consistent with the scaling theories stating that V is inversely proportional to the isentropic slope (Held and Larichev 1996). Following previous literature (e.g., Barry et al. 2002), we further diagnose the average meridional mixing length $L \propto F/VT_y$, where F is the meridional eddy heat flux, and T_y is the meridional temperature gradient. Upper- and midtropospheric

FIG. 4. Changes in zonal mean eddy momentum flux resulting from BC in the (a) tropics, (b) midlatitudes, and (c) high latitudes at $\sigma = 0.38$. The black contours denote the climatological mean from the control run. The hatching represents significance at the 0.05 level.

BC results in a 6% and 4% decrease in the eddy length scale, respectively. The decrease in the mixing length is consistent with the Rhine's scale, $L_\beta = (V/\beta)^{1/2}$, where β is the meridional gradient of the Coriolis parameter, at which the inverse energy cascade is halted by the β effect (Held and Larichev 1996; Barry et al. 2002). A detailed discussion on the scaling arguments for baroclinic eddies is beyond the scope of this paper, but we hope that similar tropospheric heating experiments in GCMs can be used to test eddy closure theories in future studies.

In this study we focus on changes in the zonal mean flow. In reality the longitudinal variations in Earth's

FIG. 5. Changes in meridional streamfunction resulting from BC at $\sigma =$ (a) 0.38, (b) 0.60, and (c) 0.90. The black contours denote the climatological mean from the control run. Positive values indicate clockwise motion and negative values indicate counterclockwise motion. The hatching represents significance at the 0.05 level.

surface and the inhomogeneous forcing can excite stationary waves and thus result in changes in the zonally asymmetric flow (e.g., Held et al. 2002; Ming et al. 2011). The effects of these two factors are always coupled with each other and may not be separated cleanly. Here we only include one source of zonal asymmetry by using an AGCM with realistic geography, but with idealized BC forcing, and the role of changes in stationary eddies may be underestimated. Future work may focus on the atmospheric response to idealized BC forcing in AGCMs with idealized boundary conditions (e.g., aquaplanet models) or to inhomogeneous BC forcing in realistic AGCMs.

FIG. 6. Changes in northward energy flux (red solid) and contributions from the mean meridional circulation (black solid), stationary eddies (black dashed), and transient eddies (black dotted) resulting from BC at $\sigma =$ (a) 0.38, (b) 0.60, and (c) 0.90. Note that the y-axis scale is different in (a), (b), and (c).

c. Energy budget

The above analysis shows that the temperature response is key to understanding the extratropical circulation change. Free-tropospheric BC affects the static stability and meridional temperature gradient at

midlatitudes, which weakens the baroclinic eddies and thus the meridional energy transport. It is clear that the temperature and circulation changes resulting from upper-tropospheric BC are much stronger than that resulting from midtropospheric BC. Boundary layer BC, in contrast, does not have a significant effect on temperature, zonal wind, or eddy activity.

The altitude dependence of BC-induced response is not immediately intuitive. Since we use the same BC burden in the three experiments, the increase in atmospheric shortwave absorption is similar and cannot explain the different magnitudes of temperature change. An analysis of the change in heating rates provides some insights into the temperature response (Fig. 8). Atmospheric temperature is affected by physical processes including radiative shortwave (SW) heating and long-wave (LW) cooling, latent heat release by convective and large-scale cloud formation, vertical diffusion, and dynamical advection of sensible heat. BC at different altitudes results in local SW heating anomalies. Note that the SW heating caused by boundary layer BC is stronger than that caused by free-tropospheric BC because model layers in the boundary layer contain less mass. The vertical-integrated SW heating actually increases slightly with altitude as a result of low and middle clouds. Since we focus on the equilibrium response to a perturbation, the changes in heating rates by different processes have to balance out one another. Therefore, as diabatic heating terms (radiative and latent heating) and vertical diffusion are computed directly from the model output, one can evaluate dynamical advection as a residual.

When BC is added in the free troposphere, the most important sources of heating rate changes are SW radiation, latent heat release by large-scale precipitation, and dynamical advection. The forced increase in SW heating in the upper troposphere is mainly offset by a decrease in dynamical heating (Fig. 8a), while the reduction in large-scale precipitation and dynamical heating are almost equally important in balancing out the stronger SW heating in the midtroposphere (Fig. 8b). Note that the change in LW radiation is small despite the strong local warming. Further analysis indicates that there is a decrease in the cloud amount at the heating layer. This leads to a decrease in LW emissivity that balances out the higher temperature, and as a result there is only a small change in LW radiation. Below the heating layer, the increase in dynamical heating contributes to a higher temperature, which is damped by decreased convective heating. The large response of dynamical advection compared to other heating sources indicates the change in the large-scale circulation is the main mechanism for reestablishing the

FIG. 7. Changes in northward energy flux by (left) transient eddy and (right) eddy kinetic energy resulting from BC at $\sigma =$ (a),(d) 0.38, (b),(e) 0.60, and (c),(f) 0.90. The black contours denote the climatological mean from the control run. The hatching represents significance at the 0.05 level.

atmospheric energy balance under a heating perturbation in the free troposphere.

The energy balance change resulting from boundary layer BC is very different. The warming at approximately 900 hPa stabilizes the boundary layer and thus suppresses turbulent diffusion of sensible heat and shallow convection. As a result, the increased SW absorption in the heating layer is mainly damped by sub-grid vertical diffusion and a decrease in convective heating. Below the heating layer, LW cooling becomes stronger and is balanced by the resulting increase in latent heat release by large-scale condensation. The change in dynamical advection is very small, indicating that boundary layer BC is less capable of altering atmospheric circulation than free-tropospheric BC. This is consistent with the result that boundary layer BC does not cause significant changes in zonal wind or baroclinic eddies.

We conclude this section by noting that the heating rate changes caused by LW radiation and latent heat release are closely related to cloud changes. While it is expected that the model-simulated extratropical responses are more robust than tropical responses, which may be strongly affected by uncertainties in convective parameterizations, we emphasize that it remains to be seen whether other GCMs may yield similar results. In addition, BC can also result in changes in clouds through the indirect effect and the cloud-absorption effect (Jacobson 2012), which are not considered here. These effects are potentially large but uncertain, and inclusion of them would likely quantitatively alter our results.

4. Theory

To further understand why the temperature response to upper-tropospheric BC is much stronger, we examine

FIG. 8. Changes in midlatitude heating rates caused by SW and LW radiation, latent heat release by convective (CV) and large-scale (LS) cloud formation, subgrid vertical diffusion (VD), and DY resulting from BC at $\sigma =$ (a) 0.38, (b) 0.60, and (c) 0.90.

how the change in dynamical advection resulting from free-tropospheric BC occurs. The advection of sensible heat DY can be divided into contributions from the mean meridional circulation and eddies (both stationary and transient):

$$\text{DY} = -\frac{\bar{v}}{a} \frac{\partial \bar{T}}{\partial \phi} - \bar{\omega} \left(\frac{\partial \bar{T}}{\partial p} - \frac{R\bar{T}}{C_p p} \right) - \frac{1}{a \cos \phi} \frac{\partial \bar{v}'T'}{\partial \phi} - \left(\frac{\partial \bar{\omega}'T'}{\partial p} - \frac{R\bar{\omega}'T'}{C_p p} \right). \quad (1)$$

Here v is the meridional wind, ω is the vertical pressure velocity, T is the temperature, a is the radius of the Earth, ϕ is the latitude, p is the pressure, R is the gas constant, and C_p is the specific heat capacity of air. Overbars denote monthly and zonal means, and primes denote deviations thereof. The first and second right-hand-side terms of Eq. (1) are the meridional and vertical advection of heat by the mean meridional circulation, respectively. The third and fourth terms are meridional and vertical eddy heat flux convergence, respectively.

Figure 9 shows the vertical profiles of changes in different terms of Eq. (1) at midlatitudes resulting from free-tropospheric BC. Note that the explicitly computed DY change agrees approximately with the inferred one in Fig. 8. It is clear that the response of dynamical advection is dominated by the change in vertical eddy heat flux convergence, which cools the heating layer and warms the atmosphere below. There is also anomalous mean advective warming associated the weaker Ferrel cell (Fig. 5) in the upper-tropospheric BC perturbation case, but the magnitude is much smaller. The change in meridional eddy heat flux convergence is also small. This is because the

strongest weakening of the meridional eddy heat flux at the jet core (not shown) leads to an increase (decrease) in the heat flux convergence on the equatorward (poleward) flank of the jet, which cancel out when averaged over the midlatitudes.

In section 3c we have shown the dominant balance between dynamical advection and SW heating. Neglecting the small terms in Eq. (1) and using potential temperature θ instead of temperature to simplify the equation, we have

$$\delta \left\langle \frac{\partial \omega' \theta'}{\partial p} \right\rangle dp \approx \langle Q \rangle, \quad (2)$$

where Q is the heating rate by BC-induced SW absorption and the angle brackets denote a horizontal average (midlatitudes in this study). Integrating from the bottom of the heating layer to the tropopause at midlatitudes and since the vertical heat flux at the tropopause is approximately zero, Eq. (2) then becomes

$$\delta \langle \bar{\omega}' \theta' \rangle_h \approx \{Q\}, \quad (3)$$

where $\{Q\} = \int_{p_t}^{p_h} \langle Q \rangle dp$ and subscripts h and t denote the bottom of the heating layer and the tropopause, respectively. Equation (3) indicates that the weaker vertical eddy heat flux across the heating level acts to balance the anomalous SW heating above it. Note that $\{Q\}$ resulting from upper- and midtropospheric BC are similar.

It is tempting to relate the change in vertical eddy heat flux to the change in static stability as our ultimate goal is to understand the temperature response. In the interior of the extratropical troposphere, the total eddy heat flux is roughly aligned along the mean isentropes (Held and Schneider 1999). In the pressure coordinate this can be written as

FIG. 9. Changes in midlatitude adiabatic heating rates averaged at midlatitudes resulting from BC at $\sigma =$ (a) 0.38 and (b) 0.60. Solid lines represent changes in the meridional (black) and vertical (red) advection of heat by the mean meridional circulation. Dashed lines represent changes in meridional (black) and vertical (red) eddy heat flux convergence.

$$-\frac{\overline{\langle \omega' \theta' \rangle}_h}{\overline{\langle v' \theta' \rangle}_h} = \frac{\langle \partial_y \bar{\theta} \rangle_h}{\langle \partial_p \bar{\theta} \rangle_h}. \quad (4)$$

The horizontal eddy heat flux can be related to the mean meridional temperature gradient using the eddy diffusivity of heat D , that is, $\langle v' \theta' \rangle_h = -D \langle \partial_y \bar{\theta} \rangle_h$. Therefore, Eq. (4) can be written as

$$\overline{\langle \omega' \theta' \rangle}_h = \frac{D \langle \partial_y \bar{\theta} \rangle_h^2}{\langle \partial_p \bar{\theta} \rangle_h}. \quad (5)$$

Since the atmosphere is perturbed by globally uniform BC at a certain level in this study, one would expect the change in meridional temperature gradient $\langle \partial_y \bar{\theta} \rangle_h$ is small. It can be seen in Fig. 1 that the temperature response at midlatitudes does not have much meridional difference at midlatitudes. More detailed calculations show that the change in $\langle \partial_y \bar{\theta} \rangle_h$ resulting from upper-tropospheric (midtropospheric) BC is less than 5% (1%) when averaged over the midlatitudes, despite some spatial variations within the midlatitudes. If the changes in D are also small, the perturbation of the vertical eddy heat flux can be approximated as

$$\delta \overline{\langle \omega' \theta' \rangle}_h \approx -D \frac{\langle \partial_y \bar{\theta} \rangle_h^2}{\langle \partial_p \bar{\theta} \rangle_h^2} \delta \langle \partial_p \bar{\theta} \rangle_h = -DI^2 \delta \langle \partial_p \bar{\theta} \rangle_h, \quad (6)$$

where $I = \langle \partial_y \bar{\theta} \rangle_h / \langle \partial_p \bar{\theta} \rangle_h$ is the isentropic slope. We use the change in bulk static stability below the heating level to approximate the stratification change at the heating level, that is, $\delta \langle \partial_p \bar{\theta} \rangle_h \approx (\delta \langle \bar{\theta} \rangle_h - \delta \langle \bar{\theta} \rangle_s) / (p_h - p_s)$, with the subscript s denoting the surface. We further neglect the change in surface temperature since SST is fixed, that is, $\delta \langle \bar{\theta} \rangle_s = 0$. Equation (6) then becomes

$$\delta \overline{\langle \omega' \theta' \rangle}_h \approx -DI^2 \frac{\delta \langle \bar{\theta} \rangle_h}{p_h - p_s}. \quad (7)$$

Combining Eqs. (3) and (7) yields

$$\delta \langle \bar{\theta} \rangle_h \approx \{Q\} D^{-1} I^{-2} (p_s - p_h). \quad (8)$$

Equation (8) shows that temperature change caused by a certain amount of heating is determined by the diffusivity, the isentropic slope, and the pressure difference between the surface and the heating level. Both the diffusivity and the isentropic slope have a vertical structure with lower values at higher altitudes (Chen and Plumb 2014), while the pressure difference is larger for a forcing at higher altitude. As a result, all three factors contribute to a stronger temperature change caused by heating in the upper troposphere.

The above analysis highlights the role of baroclinic eddies in reestablishing atmospheric energy balance at midlatitudes in the presence of BC-induced SW heating. Since the change in the vertical eddy heat flux tends to diffuse the anomalous heating away from the heating layer, one would expect the warming signal penetrates more to the lower troposphere at midlatitudes in the Northern Hemisphere where eddies are more energetic. This is clearly shown in our AGCM-simulated temperature response (Fig. 1). To confirm the importance of baroclinic eddies in driving atmospheric response at midlatitudes, we conduct similar heating experiments with the dry idealized model (section 2). Figure 10 compares the temperature changes in the AGCM and in the idealized model, which have qualitatively similar vertical profiles. The magnitude of temperature change increases with height before reaching its maximum at the heating level. The temperature change caused by

FIG. 10. Vertical profiles of temperature changes at midlatitudes resulting from (a) BC at $\sigma = 0.38$ (red) and 0.60 (green) in AM2 and (b) heating at $\sigma = 0.38$ (red) and 0.58 (green) in the idealized model.

upper-tropospheric heating is 2.4 (1.9) times of that caused by midtropospheric heating in the AGCM (idealized model). In the AGCM the shortwave absorption of BC becomes more effective as BC rises above the reflective cloud layer, and model-simulated SW heating caused by upper-tropospheric BC is larger than that caused by midtropospheric BC by approximately 20% (not shown). If taking into account the vertical variation in heating, the ratio in the idealized model would be about $1.9 \times 1.2 = 2.28$, even closer to the AGCM result. The similarities between the AGCM and the idealized model demonstrate the dominant role of dry dynamics in determining temperature response at midlatitudes to anomalous heating in the free troposphere. We also notice some differences between the two models. The maximum warming in the idealized model is larger than that in the AGCM by about a factor of 2, and the warming below the heating level is weaker in the idealized model. The discrepancies indicate the influence of other factors in the AGCM (e.g., convection, radiation, and boundary layer processes) on the thermal structure of the midlatitudes and thus the atmospheric response to BC-induced heating.

In deriving Eq. (8) we make an important assumption that the change in eddy diffusivity D is small. A constant D would allow us to avoid much discussion on the specific scaling of diffusivity and temperature gradient and simplify the equations. This may not be a strictly valid assumption since free-tropospheric BC has a significant influence on baroclinic eddies (section 3b). The above derivation, however, can be generalized to a case in which D is not a constant. Despite different forms, almost all the scaling relations for D used in the literature are inversely proportional to the n th power of the stratification (e.g., Green 1971; Held and Larichev 1996;

Zurita-Gotor and Vallis 2010; Jansen and Ferrari 2013). Thus, from Eq. (5) we have $\langle \omega' \theta' \rangle_h \propto \langle \partial_p \bar{\theta} \rangle_h^{-n-1}$, and Eq. (8) still holds except that there is an extra term that is proportional to $\langle \partial_p \bar{\theta} \rangle_h^{-n-1}$ on the right-hand side. This will not affect the qualitative conclusion that heating at higher altitudes yields a stronger temperature response.

5. Discussion and conclusions

The GFDL AM2 is used to examine the extratropical atmospheric-only response to global uniform BC forcings at different altitudes. BC direct and semidirect effects are considered in our analysis. Free-tropospheric BC-induced SW heating warms the troposphere with maximum temperature increase at the heated altitudes. For the same column burden, the temperature change resulting from upper-tropospheric BC is much stronger. The warming signal penetrates to a greater depth at midlatitudes than in the tropics. As a result, free-tropospheric BC stabilizes the midlatitude atmospheric column and weakens meridional temperature gradient on the equatorward flank of the tropospheric jet. Consistent with the thermal wind relation and the change in the eddy momentum flux, the response of the zonal-mean circulation to upper-tropospheric BC features a strong weakening and poleward shift of the jet. Midtropospheric BC weakens the jet without significantly shifting its location. Boundary layer BC yields slight warming of the troposphere and has a weak impact on the jet.

Free-tropospheric BC results in weaker mean meridional circulation and less energetic baroclinic eddies at midlatitudes. The weakening of the eddies is characterized by a smaller eddy velocity related to the stronger stratification and a shorter mixing length consistent with the Rhine's scale. The less energetic eddies result in a

reduction in the meridional energy transport by transient eddies, which dominates the change in total meridional energy transport at midlatitudes. Similar to the temperature response, the weakening of eddy activities and associated energy transport caused by upper-tropospheric BC is much stronger than that caused by midtropospheric BC. Boundary layer BC does not have a strong influence on the mean circulation and baroclinic eddies.

An investigation of changes in heating rates at midlatitudes helps explain the altitude dependence of the temperature response to BC-induced heating, which is key to understanding the response of extratropical circulation. A large fraction of the BC-induced boundary layer SW heating is damped by vertical diffusion of sensible heat. As a result, boundary layer BC only causes a small temperature change and does not effectively alter the large-scale circulation. BC-induced free-tropospheric SW heating causes a strong change in the vertical profile of dynamical heating, which is dominated by the change in vertical eddy heat flux convergence. There is a reduction in vertical eddy heat transport to the heating level, which balances the BC-induced local SW heating and warms the atmosphere below the heating layer. Upper-tropospheric BC results in a stronger temperature response since the eddy diffusivity and the isentropic slope decrease with height and the increase in stratification extends to higher altitudes. Similar results are found when using a dry idealized model, which further highlights the importance of dry dynamics in driving the temperature change at midlatitudes. Other factors, such as moisture and radiation, also affect the extratropical response, but their impacts are secondary.

The strong atmospheric-only response at midlatitudes suggests that BC is capable of altering weather patterns, as the underlying dynamics involved operate on synoptic time scales. Our results suggest that BC may also modulate extratropical cyclones and affect midlatitude extreme weather. Preliminary results (not shown) indicate that upper-tropospheric BC leads to increases in light precipitation frequency and decreases in moderate-to-heavy precipitation frequency over the storm-track regions and reduces total precipitation by approximately 20%. Midtropospheric BC yields similar but weaker changes in precipitation extremes. The decreases in midlatitude extreme precipitation caused by free-tropospheric BC is consistent with weaker baroclinic eddies. Since BC is mainly removed from the atmosphere through wet scavenging, the decrease in precipitation may further amplify the circulation response when BC concentration is interactive with atmospheric circulation rather than prescribed, as in this study.

The regional perturbation experiments suggest that the atmospheric response to BC is mostly local and linearly additive, and the extratropical response examined in this study is ascribed mainly to midlatitude BC. Thus, the results presented here have important implications for understanding the climate impacts of realistic BC, which concentrates at midlatitude industrial regions in the Northern Hemisphere. The strong altitude dependence of BC-induced response indicates that BC at higher altitudes, although less abundant, may still have large impacts on climate. This highlights the importance of better constraining the spatial distribution of BC concentration, which is currently uncertain across global models and observations (Koch et al. 2009; Bond et al. 2013). In general, model-simulated BC has a high bias in the upper troposphere (Allen and Landuyt 2014), and therefore it is likely that the BC-induced atmospheric responses presented here are overestimated in current models.

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