Circulation and Tides in a Cooler Upper Atmosphere: Dynamical Effects of CO₂ Doubling

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Abstract Thermosphere cooling is a known effect of increasing CO₂ in the atmosphere. In this study, we explore the changes of thermosphere circulation and tides in the cooled thermosphere via a doubled CO₂ numerical experiment using the Ground-to-topside Atmosphere Ionosphere model for Aeronomy (GAIA). The results reveal three major features. (1) The thermosphere cools about 10 K more around solstices than equinoxes, more at the summer pole than the winter pole. (2) The meridional circulation shifts downward and strongly accelerates by 5–15 m s⁻¹. (3) The tidal activity experiences dramatic changes, with a 40–60% reduction in the semidiurnal tides (SW2) throughout the thermosphere but an 30–50% enhancement in diurnal tides (DW1) below 200 km altitude. The nonmigrating tide DE3 has only minor changes. These changes in temperature, meridional circulation, and tides are persistent features in all seasons and can profoundly affect the spatial distribution and diurnal cycles of the ionospheric responses to CO₂ doubling via atmosphere composition and electrodynamics.

1. Introduction

Global cooling is the impact of increasing CO₂ concentration on the middle and upper atmosphere, in contrast to global warming in the troposphere. Upper atmosphere cooling, or thermosphere cooling, was first predicted by Roble and Dickinson (1989) using the 1-D globally averaged National Center for Atmospheric Research-Thermosphere Ionosphere General Circulation Model (NCAR-TIGCM). They found that the cooling can reach about 50–60 K with a doubling of the CO₂ mixing ratio from about 350 to 700 ppm. This cooling leads to a thermosphere density drop of ~30–60% at heights above 200 km. Later, numerical simulations using more sophisticated 3-D NCAR models, for example, Thermosphere Ionosphere General Circulation Model (TIGCM), Thermosphere Ionosphere MesoSphere General Circulation Model (TIMEGCM), and Whole Atmosphere Community Climate Model-eXtended (WACCM-X), all found consistent results with this early prediction (see, e.g., Qian et al., 2011; Rishbeth & Roble, 1992; Solomon et al., 2018). A Spectral Mesosphere/Lower Thermosphere Model (SMLTM, upper boundary at 220 km) was used by Akmaev and Fomichev (1998), who also found a cooling of ~50 K with a density reduction of about 45% near 200 km altitude in response to a CO₂ doubling from 360 to 720 ppm. These predicted cooling effects have been confirmed by direct or indirect observational evidences from long-term measurements of satellite drag (e.g., Emmert et al., 2010) and some ionospheric parameters like the height of the maximum electron density of the F2 region (hmF2) and ion temperature (Laštovička et al., 2012; Ogawa et al., 2014; Zhang et al., 2016). Thus, thermosphere cooling as the thermal response to increasing CO₂ concentration is well established by now.

On the other hand, dynamical responses (e.g., the mean circulation and tides) in the thermosphere remain unclear. Since thermospheric circulation and tides are key aspects of the thermosphere-ionosphere system, we naturally ask how will the mean circulation and tidal activities change in a cooler thermosphere, and how do they contribute to the ionospheric response? As a first step to answer these questions, we carried out a doubled CO₂ numerical experiment using the whole atmosphere model Ground-to-topside Atmosphere Ionosphere model for Aeronomy (GAIA).

2. The GAIA Model and Experimental Setup

GAIA is a 3-D self-consistent, fully coupled whole atmosphere model of the Earth’s troposphere, stratosphere, mesosphere, thermosphere, and ionosphere, covering the altitude range from the ground to ~600 km.
for the neutrals and to 3,000 km for the plasma (Jin et al., 2012). It has a horizontal resolution of 2.8° × 2.8° (latitude × longitude) and a vertical resolution of 0.2 scale height. The model cannot resolve gravity waves (GWs). Instead, it uses parameterizations to account for GWs, with formulations by McFarlane (1987) for orographic GWs and those by Lindzen (1981) for nonorographic GWs. In the troposphere, stratosphere, and mesosphere, a full radiation scheme developed by Nakajima et al. (2000) is used. The distribution of O₃ is climatologically prescribed. In the thermosphere, the radiative cooling parameterization by Fomichev et al. (1999) is used to calculate the cooling rates by CO₂. The GAIA radiative scheme has a flexibility to use different amounts of CO₂ in the range of 150–720 ppm.

GAIA has been demonstrated to be a powerful model that captures comprehensive coupling processes between the neutral and ionized atmosphere, successfully reproducing prominent lower atmosphere-driven features in the thermosphere and ionosphere, for example, the Wave-4 structure, the thermosphere cooling during stratosphere sudden warmings, and El Niño–Southern Oscillation (ENSO) signatures in the upper atmosphere (Jin et al., 2011; Liu et al., 2009, 2014, 2017).

To examine effects of doubled CO₂ on thermospheric dynamics, we performed two GAIA simulations with CO₂ concentration of 345 ppm (corresponding to the value around 1985) and 690 ppm, respectively. The value of 345 ppm, instead of the current value of ∼415 ppm, is chosen to better compare with previous CO₂ doubling studies (Akmaev & Fomichev, 1998; Roble & Dickinson, 1989) that used similar values. The model was run for 2 years (24 months) under solar minimum (F10.7 = 80 sfu) and geomagnetically quiet conditions (cross-polar cap potential = 30 kV). Simulation results in the latter year from January to December are used to examine the doubled CO₂ effects, which are expressed as the difference between the two model runs (2×CO₂-CO₂). In this brief report, we focus on the thermosphere response, while the ionosphere response will be reported separately.

3. Results

In this section, we first examine the double CO₂ impact on the thermosphere temperature and compare it with published results for consistency checks and then examine changes in the dynamics and tidal activities.

3.1. Thermal Responses

Figure 1 presents the impacts on zonal mean thermosphere temperature (ΔTₙ) and mass density (Δρ) in June, with the left panels for globally averaged 1-D height profiles and the right for the height-latitude distribution.

When averaged globally (see the left column of Figure 1), ΔTₙ shows thermosphere cooling up to −60 K above 200 km height, along with a density reduction (Δρ) of 40–60%. Positive ΔTₙ of 5–10 K occurs between about 110 and 150 km altitudes, coinciding with a minimum in Δρ. When unfolded in latitude (see the upper right panel), we notice that the thermosphere cooling is latitude dependent, being stronger in tropics (∼−70 K) than polar regions, and stronger near the summer pole (∼−50 K) than the winter one (∼−30 K). This latitudinal asymmetry is likely a result of altered meridional circulation described in the next section. The positive ΔTₙ at lower thermosphere (110–150 km), on the other hand, is mainly confined to polar regions. We note here that this positive ΔTₙ does not indicate any heating source. Rather, it is largely an “apparent” heating caused by the combined effect of thermal contraction (lowering of the pressure surfaces) and the vertical temperature gradient (Akmaev & Fomichev, 1998; Rishbeth & Roble, 1992).

The above good agreement in the thermal responses of thermosphere to CO₂ doubling between GAIA results and previous NCAR-GCMs (e.g., Roble & Dickinson, 1989) and SMLTM (Akmaev & Fomichev, 1998) studies demonstrates consistency between these physics-based models to simulate the magnitude and vertical structures of the temperature changes above 100 km.

In addition to latitudinal and height structures shown above, we further examined the day-night and seasonal variations of the thermal response. As seen in the left panel of Figure 2, the thermosphere cooling shows a day-night difference (∼5–10 K), with stronger cooling at night caused by larger downward heat conduction. Meanwhile, the globally averaged ΔTₙ in the right panel exhibits a clear seasonal variation. For instance, the thermosphere cooling above 300 km is ∼−60 K around solstices and ∼−50 K around equinoxes, demonstrating stronger thermal impacts around solstices.
Figure 1. Responses to doubled CO₂ in temperature (ΔTn) and neutral mass density (Δρ in percentage) in June. (left column) Global average values. (right column) Latitude-height distribution.

3.2. Dynamical Responses

Figure 3 shows the zonal mean zonal wind (U), meridional wind (V), and vertical wind (W) around June solstice and September equinox for base CO₂, along with their responses (ΔU, ΔV, and ΔW) to doubled CO₂. Near June solstice, changes in the zonal wind (ΔU) are eastward (positive) in most regions except for mid-latitudes in the lower thermosphere (100–150 km). This consequently accelerates the core of the eastward wind jet in the winter (southern) hemisphere by 5–15 m s⁻¹, and the westward wind in the summer (northern) hemisphere below 150 km by similar magnitude. This perturbation pattern in the zonal wind is consistent with those shown in Rishbeth and Roble (1992) for December case in terms of local season (upper panel in their Figure 6), except that their wind change is of somewhat smaller magnitude (<6 m s⁻¹).

Figure 2. (left) Height profile of ΔTn in June averaged over daytime (06–18 LT, solid line) and nighttime (18–06 LT, dashed line). (right) Height-month cross section of the globally averaged ΔTn. The pink line indicates temperature at 100 km height. It is seen that the upper thermosphere cools more around solstices than equinoxes.
Figure 3. Thermosphere mean winds (in unit of m s$^{-1}$) under base CO$_2$ conditions (first and third rows) and their responses to doubled CO$_2$ (second and fourth rows) around June solstice and September equinox. (left column) Zonal wind (U) and its change ($\Delta U$), positive eastward; (middle column) meridional wind (V) and its change ($\Delta V$), positive northward; (right column) vertical wind (W) and its change ($\Delta W$), positive upward. The white arrows depict wind directions. An acceleration of the meridional circulation is evident under doubled CO$_2$ conditions in both seasons.

Changes in the meridional circulation ($\Delta V$ and $\Delta W$) around June solstice are shown in the middle and right panels in the second row of Figure 3. We see that perturbations in both meridional ($\Delta V$) and vertical ($\Delta W$) winds are in the same direction as the original winds (indicated by the white arrows), suggesting a strengthening of the meridional circulation. For instance, the southward summer-to-winter flow is strengthened by 5–15 m s$^{-1}$ ($\Delta V$ in the second row). This is accompanied by enhanced upward flow ($\Delta W$) at the summer (north) pole and downward flow at the winter (south) pole at the order of 0.2 m s$^{-1}$. In addition to the acceleration, we can also discern a downward shift of the circulation pattern from the stronger changes at lower altitudes (most clearly shown in $\Delta V$). This downward shift appears to be a natural consequence of the thermosphere contraction.
Figure 4. Changes in three major tidal components around June solstice and September equinox. It shows significant enhancement of DW1 (10–20 K) in the lower thermosphere and sharp reduction of SW2 (10–30 K) throughout the thermosphere. DE3 shows small changes within 3 K (note different color bars).

Around September equinox, the zonal wind perturbation $\Delta U$ is predominantly westward (negative) of 5–10 m s$^{-1}$ (see the left panel in the fourth row of Figure 3). This leads to acceleration of the original westward wind at middle latitudes but deceleration of the eastward wind in polar regions. The meridional circulation near equinoxes consists of one cell in each hemisphere, which is anticlockwise in Northern Hemisphere but clockwise in Southern Hemisphere (see white arrows in V and W in the third row). Both cells are enhanced under doubled CO$_2$ conditions, as indicated by the same direction in the perturbation winds ($\Delta V$ and $\Delta W$) and the original wind (V and W) except for at South Pole above 200 km (see bottom row of Figure 3). A downward shift of the circulation patterns is also discernible. Thus, similar to June solstice, the meridional circulation is enhanced and shifted downward in the cooled upper atmosphere.

### 3.3. Tidal Responses

Figure 4 displays the changes in tidal amplitudes due to CO$_2$ doubling for three major components, that is, the migrating diurnal component DW1 and semidiurnal component SW2, and the nonmigrating DE3 component. The DW1 shows a strong enhancement of 10–20 K (30–50%) between about 150 and 200 km at low and middle latitudes, with some extension to higher altitude at middle latitudes. In contrast, a large reduction occurs in SW2 throughout the whole thermosphere. Its amplitude is reduced by 10–25 K, which corresponds to 40–60% at middle latitudes. Though somewhat smaller in magnitude, migrating tidal amplitude changes in DW1 and SW2 show similar structures around September equinox (bottom row of Figure 4), indicating the persistence of these changes throughout the year.
Figure 5. Local time-latitude variation of $\Delta T_n$ at 400 km altitude and the maximum $F$ region height $\Delta hmF_2$. Though being all negative values, a quasi-semidiurnal pattern is clearly visible at low and middle latitudes.

On the other hand, the nonmigrating tide DE3 experiences small changes below 3 K (~20%), whose sign varies with season. Thus, the impact on DE3 appears to be minor, though its robustness may need further examination due to its small magnitude.

The large change in SW2 can have significant consequences on the local time dependence of the thermosphere/ionosphere responses. As an example, Figure 5 shows the perturbation in temperature at 400 km and the maximum ionosphere height ($hmF_2$) at different local times. A quasi-semi diurnal pattern is evident in both parameters, demonstrating the tidal modulations. More comprehensive analysis of the ionospheric responses will be carried out in a separate study.

4. Discussions

We have examined the impacts of doubled CO$_2$ concentration on the thermal and dynamical structures of the upper atmosphere using the GAIA model. The analysis confirms the thermosphere cooling reported in the literature in terms of globally averaged thermosphere temperature and mass density responses. It further reveals three pronounced new features in the cooled thermosphere: (1) larger thermosphere cooling around solstices than equinoxes, (2) faster meridional circulation, and (3) strong impact on migrating tides. We briefly discuss these features in the following.

4.1. Stronger Thermosphere Cooling Around Solstices

The thermosphere cooling is larger around solstices than around equinoxes. Above 300 km, it is about $-60$ K around solstice and $-50$ K around equinoxes (see right panel of Figure 2). One possible cause for this could be the seasonal variation of the temperature in the lower thermosphere. As indicated by the pink line in Figure 2, the temperature at 100 km is higher around solstices than equinoxes. Since the direct radiative cooling rate by CO$_2$ in the lower thermosphere has a strong positive temperature dependence (Mlynczak et al., 2010), higher temperature around solstice leads to a higher cooling rate there. This would consequently lead to faster downward thermal conduction, hence stronger upper thermosphere cooling around solstices.

4.2. Intensification of the Meridional Circulation

The meridional circulation in the thermosphere accelerates in response to CO$_2$ doubling. This result is somewhat counterintuitive, as we would generally expect weaker dynamics when the thermosphere loses both its thermal and potential energy via cooling and shrinking. However, similar responses with a faster meridional circulation were also found in the middle atmosphere by Rind et al. (1990). They attributed the acceleration to increased eddy forcing and GWs caused by enhanced instability, which is in turn due to the combination of troposphere warming and stratosphere cooling. Though their theory cannot be directly applied here, we note that the lower thermosphere warming and upper thermosphere cooling produces a similar situation in the vertical temperature profile, which may potentially enhance the GWs in the thermosphere. Actually, the increase of GW activities in the mesosphere and lower thermosphere (MLT) region and the thermosphere in response to increasing CO$_2$ has been suggested by model simulations (Akmaev & Fomichev, 1998).
and ionospheric observations (Oliver et al., 2013), respectively. Thus, it seems reasonable to speculate that the intensification of the thermospheric meridional circulation is likely caused by enhanced GW activities via the influence of GW drag, in analogy to mechanisms in the middle atmosphere discussed by Rind et al. (1990) and Akmaev and Fomichev (1998).

Regardless of its causes, the stronger meridional circulation has an apparent impact on the thermal response. In particular, vertical motions near the poles (upward at the summer pole and downward at the winter pole) may have likely contributed to the hemispheric asymmetry of the thermosphere cooling (see top right panel of Figure 1) via their adiabatic effect. This hemispheric asymmetry with stronger cooling near the summer pole was also seen in results of SMLTM (Akmaev & Fomichev, 1998) and NCAR-GCMs (Rishbeth & Roble, 1992) but with no explanation.

4.3. Strong Impacts on Migrating Tides

The DW1 increases by 10–20 K (30–50%) in the lower thermosphere, while SW2 decreases by 10–25 K (40–60%) throughout the whole thermosphere. The increase in DW1 is also reflected in the larger day-night difference in the lower thermosphere (see left panel in Figure 2). Changes in DW1 above 150 km can be related to changes in (a) tidal forcing by solar heating absorbed by O2 and N2 and (b) tidal dissipation by molecular viscosity and heat conductivity. In a cooler thermosphere, tidal forcing drops due to lower concentrations of O2 and N2, and tidal dissipation weakens due to smaller molecular viscosity and heat conduction. The enhancement of DW1 in response to CO2 doubling seems to suggest that the drop of tidal forcing is overcompensated by the weakening of tidal dissipation.

On the other hand, SW2 in the thermosphere comes predominantly from the lower atmosphere (Forbes, 1995). Its variability indicates changes in either the generation source (mainly ozone in the stratosphere) or the propagation condition in the middle atmosphere. Since ozone remains unchanged under doubled CO2 conditions in our model simulation, the SW2 reduction has to come from the latter. Figure 6 shows the zonal wind perturbation in the middle atmosphere, which is dominantly westward of 5–10 m s−1 around both equinoxes and solstices. This westward perturbation, via the wind filtering effect, would cause the observed weakening of the SW2 (a westward propagating wave). We note that neglecting ozone change might potentially affect the SW2 variation, as cooling in the stratosphere could theoretically lead to ozone enhancement, hence increasing SW2 tidal forcing. An accurate assessment of this feedback effect requires interactive chemistry that GAIA does not include at this moment.

5. Conclusion

The GAIA experiment reveals that doubling CO2 not only cools the upper atmosphere but also significantly disturbs its dynamics. In particular, the meridional circulation is strongly accelerated by 5–15 m s−1, while the semidiurnal tide SW2 is sharply reduced by 40–60%. It would be interesting for other models and observations to verify these predictions. Comparing different models could help elucidate the role of various processes in global changes of the upper atmosphere. Though thermosphere observations are sparse, the tidal response can be relatively easily investigated using the more amply available ionospheric observations. Dynamical and tidal changes can profoundly affect the redistribution of thermosphere composition and ionosphere electrodynamics, hence should be taken into account in any assessment of double CO2 effects on the upper atmosphere.

Data Availability Statement

The data used in this study are publicly available online (https://doi.org/10.5281/zenodo.3653907).

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Figure 6. Zonal wind and its perturbation in the middle atmosphere between 50 and 100 km around June solstice and September equinox (in unit of m s$^{-1}$). We see that the perturbation wind is dominantly westward in both seasons.

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