Mechanisms for Zonal Mean Wind Responses in the Thermosphere to Doubled CO$_2$ Concentration

Masaru Kogure$^1$, Huixin Liu$^1$, and Chihiro Tao$^2$

1Department of Earth and Planetary Science, Kyushu University, Fukuoka, Japan, 2National Institute of Information and Communications Technology, Tokyo, Japan

Abstract We explored mechanisms for the thermospheric zonal mean wind responses to doubled CO$_2$ concentration through investigating the zonal mean momentum balance in the thermosphere using GAIA model simulations in June. The analysis shows that ion drag, molecular viscosity, and meridional pressure gradient force vary 3–20 times more than the other forces due to the doubled CO$_2$ concentration. The three forces strongly attenuate each other; consequently, the increase in zonal mean zonal ion drag dominantly strengthens the southward wind ($\sim$15 ms$^{-1}$ at maximum) in the northern (summer) hemisphere and latitudes north of 35 °S. This strengthened ion drag was attributed to increased ion density/relative velocity between ions and neutral particles in the northern/southern hemisphere. Southward of 35°S, the zonal pressure gradient force and the meridional advection of zonal wind strengthen the southward wind by 4–12 ms$^{-1}$ in total. On the other hand, the zonal wind is mainly altered by increased meridional pressure gradient force ($\sim$15 ms$^{-1}$ at maximum) caused by a latitudinally asymmetric response of the thermosphere density to increasing CO$_2$ concentration. Our results suggest that the increased ion density in the northern hemisphere is the underlying trigger for the wind responses to increasing CO$_2$ concentration. Furthermore, we show that the meridional advection was mainly ($\sim$70%) due to that of the DW1 tidal component.

1. Introduction

The thermosphere is cooling with increasing CO$_2$ concentration, while the troposphere is warming. This cooling trend in the thermosphere has been predicted by several numerical models, and the models also show that this trend leads to a thermospheric density drop (Cnossen, 2020; H. Liu et al., 2020; H. Liu et al., 2021; Qian et al., 2011; Rishbeth & Roble, 1992; Roble & Dickinson, 1989; Solomon et al., 2018). The ionosphere is cooling, as well as the thermosphere, and the peak altitude of the electron density (i.e., h$_m$F$_2$) descends and the density increases (decreases) below (above) $\sim$230–300 km (Laštovička et al., 2006, 2008; Qian et al., 2009). These trends have been confirmed by satellite observations (e.g., Emmert, Fejer, et al., 2004; Emmert, Picone, et al., 2004; Keating et al., 2000; Marcos et al., 2005) and incoherent scatter radars (e.g., Ogawa et al., 2014; Zhang & Holt, 2011).

Although the thermal cooling due to increased CO$_2$ concentration is understood well, the dynamical response in the thermosphere remains unclear. According to Laštovička (2021), Brum et al. (2012) are the only published observation study on long-term trends above 150 km altitude until now. Brum et al. (2012) used Fabry-Perot nightglow emission measurements over Arecibo during 1980–2010 and reported a strong dependency of the wind trends on the local time and season. Although such a long-term observation is important to assess the CO$_2$ effect on the real thermosphere, it has a limitation for understanding the mechanism and global trend. Recently, H. Liu et al. (2020) explored doubled CO$_2$ concentration effects on the thermospheric circulation and tides using model simulations with the whole atmosphere model GAIA (Ground-to-topside Atmosphere Ionosphere model for Aeronomy; Jin et al., 2011). The simulation results revealed a strengthening of the zonal mean meridional wind by 5–15 m s$^{-1}$, along with a strengthening of the zonal mean zonal wind in the lower thermosphere below $\sim$150 km altitude at a solstice. This strengthening of the zonal mean wind potentially alters the steady-state global morphology of ion flow and density. Also, these dynamical changes in the zonal mean winds can strongly affect the transportation of mass and propagation of atmospheric waves (Chapman & Lindzen, 1970; Forbes, 2007; Lieberman et al., 2013; Nappo, 2012). H. Liu et al. (2020) indeed showed significant changes in tidal waves, with a 30%–50% increase in the DW1 component and 40%–60% decrease in the SW2 component. H. Liu et al. (2020) also speculated that the strengthening of the meridional circulation causes a latitudinal dependency of variations in the density and temperature in the thermosphere.
The purpose of this study is to explore the detailed underlying mechanisms for the strengthening of the zonal mean winds in terms of momentum balance (e.g., Andrews et al., 1987; Maeda et al., 1999; Tepley et al., 2011; Tsuda et al., 2007). In the thermosphere, neutral winds are mainly controlled by the balance between the pressure gradient force, ion drag, molecular viscosity, and the Coriolis force. Momentum advections by waves also have an important role in the meridional circulation although they are smaller than the main four forces (Becker & Vadas, 2020; Miyoshi et al., 2015; Vadas, 2007; Vadas & Fritts, 2004). We investigate the contributions of these terms to changes in the zonal mean wind using the same model simulations employed by H. Liu et al. (2020).

2. The GAIA Simulations With CO₂ Concentrations of 345 and 690 ppm

GAIA is a whole atmosphere-ionosphere self-coupled model and simulates the neutral atmosphere from the ground to ~600 km (exobase) and the ionosphere up to 3,000 km (Jin et al., 2011). This model couples three models: a general circulation model (GCM), an ionospheric model, and an electrodynamics model. The thermosphere is simulated by the GCM with T42L150, corresponding to a 2.8° × 2.8° horizontal, 10 km in height, and 1 hr in time. GAIA simulations with 345 and 690 ppm CO₂ concentration for 2 years under solar minimum (F10.7 = 80 sfu) and geomagnetically quiet conditions (cross-polar cap potential = 30 kV). Because of high sensitivity in high latitudes to geomagnetic activity (Forbes, 2007; H.-L. Liu et al., 2021; Solomon et al., 2015), this paper focuses on latitudes lower than 60°N/S. Since June solstice has the strongest response to doubled CO₂ concentration (H. Liu et al., 2020), this study focuses on the analysis of simulation results in June of the second year. The output data used in the following have resolutions of 2.5° × 2° (lat × lon) in horizontal, 10 km in height, and 1 hr in time.

3. Momentum Balance in the Thermosphere

To analyze changes in the momentum due to the doubled CO₂ concentration, we use the momentum equation in spherical and z-coordinates based on Maeda et al. (1999), Tsuda et al. (2007), Tepley et al. (2011), and Andrews et al. (1987). The z-coordinate (height) is preferred over the pressure coordinate for ionosphere/thermosphere studies not only because it facilitates comparison with observations (mostly made at certain heights), but also because the ionosphere and the ion drag (the most important factor in the upper atmosphere) are better organized with height. Many studies on the long-term trend in the ionosphere/thermosphere have been carried out in z-coordinate, for example, Keating et al. (2000), Emmert, Picone, et al. (2004), Emmert, Fejer, et al. (2004), Marcos et al. (2005), Saunders et al. (2011), Solomon et al. (2015), Cnossen (2020), H. Liu et al. (2020), and H.-L. Liu et al. (2021).

The momentum equation in the thermosphere can be written:

\[
\frac{\partial u}{\partial t} = -\frac{1}{\rho} \frac{1}{r \cos \theta} \frac{\partial P}{\partial \lambda} + f v + \frac{u \tan \theta}{r} + \left( \frac{J \times B}{\rho} \right)_{z} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \mu \frac{\partial u}{\partial z} \right) - \frac{u}{r \cos \theta} \frac{\partial u}{\partial \theta} - \frac{v}{r} \frac{\partial u}{\partial \theta} - w \frac{\partial u}{\partial z} + X. \tag{1}
\]

\[
\frac{\partial v}{\partial t} = -\frac{1}{\rho} \frac{1}{r \cos \theta} \frac{\partial P}{\partial \theta} - f u - \frac{u ^2 \tan \theta}{r} + \left( \frac{J \times B}{\rho} \right)_{\lambda} + \frac{1}{\rho} \frac{\partial}{\partial \lambda} \left( \mu \frac{\partial v}{\partial \lambda} \right) - \frac{u}{r \cos \theta} \frac{\partial v}{\partial \theta} - \frac{v}{r} \frac{\partial v}{\partial \theta} - w \frac{\partial v}{\partial z} + Y. \tag{2}
\]

Table 1 shows the description for each parameter. Equations 1 and 2 describe the zonal and meridional momentum balances, respectively. The term on the left hand is a temporal change of wind in the Eulerian system. The first, second, and third terms on the right hand are the pressure gradient, the Coriolis force, and geometry terms, respectively. The fourth and fifth terms are the Lorentz force (equivalent to ion drag) and molecular viscosity, respectively. \( \mu \) was calculated in accordance with Banks and Kockarts (1973) and H.-L. Liu et al. (2010). The sixth, seventh, and eighth terms are zonal, meridional, and vertical advections of wind, respectively. These wind advections correspond to momentum advections (Holton & Hakim, 2013). The terms X and Y in the momentum equations are defined as the left term subtracted from all right terms without X and Y, respectively, based on Sato et al. (2018) and Miyoshi et al. (2015). X and Y are attributed to vertical eddy viscosity that depends on the Richardson number (Miyahara et al., 1993) and artificial hyper diffusion that depends on models, hence representing model uncertainty.
Table 1

Description of Symbols for the Momentum Equation

| Symbols | Description |
|---------|-------------|
| \( u \)  | Zonal wind \([\text{m s}^{-1}]\) |
| \( v \)  | Meridional wind \([\text{m s}^{-1}]\) |
| \( w \)  | Vertical wind \([\text{m s}^{-1}]\) |
| \( \rho \) | Neutral mass density \([\text{kg m}^{-3}]\) |
| \( p \)  | Pressure \([\text{Pa}]\) |
| \( J \)  | Current density \([\text{A m}^{-2}]\) |
| \( B \)  | Magnetic field \([\text{kg m}^{-2} \text{A}^{-1}]\) |
| \( \mu \) | Viscosity coefficient \([\text{kg m}^{-1} \text{s}^{-1}]\) |
| \( f \)  | Coriolis parameter \([\text{s}^{-1}]\) |
| \( t \)  | Time \([\text{s}]\) |
| \( \lambda \) | Longitude \([\text{rad}]\) |
| \( \theta \) | Latitude \([\text{rad}]\) |
| \( z \)  | Altitude \([\text{m}]\) |
| \( r \)  | Earth’s radius \((= 6,378.137 \text{ km}; \text{Moritz, 2000})\) |
| \( X \)  | Zonal vertical eddy viscosity and artificial hyper diffusion defined as left term in Equation 1 subtracted from all right terms without \( X \) \([\text{m s}^{-2}]\) |
| \( Y \)  | Meridional vertical eddy viscosity and artificial hyper diffusion defined as left term in Equation 2 subtracted from all right terms without \( Y \) \([\text{m s}^{-2}]\) |

Applying the geographical zonal averaging to the force terms and the perturbation expansion to the geometry and advection terms in accordance with Holton and Hakim (2013), we obtained the following equations:

\[
\frac{\partial \bar{u}}{\partial t} = f \bar{u} + \frac{\mathbf{J} \times \mathbf{B}}{\rho} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \mu \frac{\partial \bar{u}}{\partial z} \right) - \frac{1}{\rho \cos \theta} \frac{\partial P}{\partial \lambda} \frac{\partial \bar{u} \tan \theta}{\partial r} + \bar{u} \frac{\partial \nu \tan \theta}{\partial r} + \bar{w} \frac{\partial \nu \tan \theta}{\partial \theta} + \bar{X},
\]

(3)

\[
\frac{\partial \bar{v}}{\partial t} = -f \bar{v} - \frac{1}{\rho} \frac{\partial P}{\partial \theta} + \frac{\mathbf{J} \times \mathbf{B}}{\rho} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \mu \frac{\partial \bar{v}}{\partial z} \right) - \frac{1}{\rho \cos \theta} \frac{\partial P}{\partial \lambda} \frac{\partial \bar{v} \tan \theta}{\partial r} - \frac{\partial \nu \tan \theta}{\partial r} - \bar{u} \frac{\partial \nu \tan \theta}{\partial r} - \bar{w} \frac{\partial \nu \tan \theta}{\partial \theta} + \bar{Y}.
\]

(4)

Here, we used the fact that \( \frac{\partial P}{\partial z} = 0 \). The overbar and prime denote the zonal average and zonally varying disturbances (i.e., wave perturbations). Since \( \frac{\partial P}{\partial z} = 0 \), the zonal mean zonal pressure gradient force arises from the zonally varying pressure and density disturbances. We calculated the monthly mean for each term in June and found out that \( \frac{\partial \bar{u}}{\partial \theta}, \frac{\partial \bar{v}}{\partial \theta}, \frac{\partial \bar{w}}{\partial \theta}, \frac{\partial \bar{u}}{\partial z}, \frac{\partial \bar{v}}{\partial z}, \frac{\partial \bar{w}}{\partial z} \), and \( \bar{w} \bar{w} \) are 10 times smaller than the other factors at least and hence can be neglected (not shown). Because the time changes of zonal mean wind are close to 0 within a month (shown in Figure S1 in Supporting Information S1), the monthly mean zonal mean flow can be regarded as the steady-state flow. Equations 3 and 4 can be simplified as follows:

\[
-f \bar{u} - \bar{w} \frac{\partial \bar{u} \tan \theta}{\partial r} + \frac{1}{\rho} \frac{\partial}{\partial \theta} \left( \mu \frac{\partial \bar{u}}{\partial z} \right) = \frac{\mathbf{J} \times \mathbf{B}}{\rho} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \mu \frac{\partial \bar{u}}{\partial z} \right) - \frac{1}{\rho \cos \theta} \frac{\partial P}{\partial \lambda} \frac{\partial \bar{u} \tan \theta}{\partial r} - \nu \frac{\partial \nu \tan \theta}{\partial r} - \frac{\partial \nu \tan \theta}{\partial \theta} - \bar{u} \frac{\partial \nu \tan \theta}{\partial \theta} - \frac{\partial \nu \tan \theta}{\partial \theta} + \bar{X},
\]

(5)

\[
f \bar{u} + \bar{w} \frac{\partial \bar{u} \tan \theta}{\partial r} = -\frac{1}{\rho} \frac{\partial P}{\partial \theta} + \frac{\mathbf{J} \times \mathbf{B}}{\rho} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \mu \frac{\partial \bar{u}}{\partial z} \right) - \nu \frac{\partial \nu \tan \theta}{\partial r} - \frac{\partial \nu \tan \theta}{\partial \theta} - \bar{u} \frac{\partial \nu \tan \theta}{\partial \theta} - \frac{\partial \nu \tan \theta}{\partial \theta} + \bar{X}.
\]

(6)

The terms in the left hand are inertial forces, and the ones in the right hand are the non-inertial forces. It should be noted that the geometry terms by waves can be regarded as a wave stress (McLandress et al., 2006). We noticed that Equation 6 is a quadratic equation for \( \bar{u} \), making it complicated to understand the relation between
the momentum and wind. Fortunately, we can neglect the geometric terms by background wind (\(A \cdot u v \tan \theta / r\)) and (\(A \cdot u^2 \tan \theta / r\)), based on the fact that the Coriolis forces account for \(\sim 90\%\) of the total inertial force in most regions of the thermosphere at middle and low latitudes below 60°S/N. Consequently, Equations 5 and 6 can be solved for \(A \cdot v\) and \(A \cdot u\), respectively as the following:

\[
\begin{align*}
\bar{v} &= f - \frac{1}{\rho \lambda} \left[ \frac{J \times B}{\rho} \right]_x + \frac{1}{\rho \gamma} \left( \frac{\partial}{\partial \theta} \left( \frac{1}{\rho} \frac{\partial u}{\partial \theta} \right) - \frac{1}{r} \frac{\partial u}{\partial \phi} \frac{\partial \rho}{\partial \phi} - \frac{1}{r} \frac{\partial u}{\partial \phi} \frac{\partial \rho}{\partial \phi} - \frac{1}{r} \frac{\partial u^2}{\partial \phi} - \frac{1}{r} \frac{\partial u \cdot v \tan \theta}{r} + \vec{X} \right] \\
\bar{u} &= f \left[ -\frac{1}{\rho} \frac{\partial \rho}{\partial \phi} \frac{\partial}{\partial \phi} + \left( \frac{J \times B}{\rho} \right) \right]_\gamma + \frac{1}{\rho} \frac{\partial}{\partial \lambda} \left( \frac{\partial v}{\partial \lambda} \right) - \frac{1}{r} \frac{\partial v}{\partial \phi} \frac{\partial \rho}{\partial \phi} - \frac{1}{r} \frac{\partial v}{\partial \phi} \frac{\partial \rho}{\partial \phi} - \frac{1}{r} \frac{\partial v^2}{\partial \phi} - \frac{1}{r} \frac{\partial v \cdot u \tan \theta}{r} + \vec{Y} \right].
\end{align*}
\]

These equations show that, due to the 90° rotation of the Coriolis force from the wind flow, the zonal mean zonal and meridional winds under steady state are determined by the meridional and zonal momentum balances, respectively. Note that \(f\) is close to 0 in the equatorial region and these equations cannot be applied there. In the following, we calculate each term on the right-hand side and examine how it contributes to changes in the zonal mean winds in responding to a doubling of CO\(_2\) concentration from 345 to 690 ppm. Those forces before the division are shown in Figures S2 and S3 in Supporting Information S1.

4. Results

4.1. Mechanisms for the Strengthening of the Zonal Mean Meridional Wind

It is well known that the ion drag, molecular viscosity, and pressure gradient are dominant forces in the thermosphere. In terms of zonal mean values, the zonal pressure gradient force is expected to be much smaller than the ion drag and molecular viscosity (first and second terms on the right-hand side of Equation 7). The ion drag and molecular viscosity are indeed 5–10 times larger than the pressure gradient and the other forces as will be seen in the following figures. Figure 1 shows the zonal ion drag and molecular viscosity divided by the zonal mean absolute vorticity (i.e., \(-f + \frac{\omega}{\rho r}\)), along with their sum and the zonal mean meridional wind averaged in June.

![Figure 1](image.png)

**Figure 1.** Zonal mean zonal forces divided by zonal mean absolute vorticity in the 345 ppm CO\(_2\) concentration experiment in June (upper row) and their responses to doubled CO\(_2\) concentration (lower row). (a) and (e) show ion drag; (b) and (f) show molecular viscosity; (c) and (g) show ion drag + molecular viscosity; (d) and (h) show zonal mean meridional wind. Positive is northward. The white contours in (a)–(d) indicate a region where the ratio of the geometry term to Coriolis force (\(A \cdot u v \tan \theta / r\)) is larger than 10% under the base CO\(_2\) concentration. The white contours in (e)–(h) are the same as (a)–(d) but under either the base CO\(_2\) or doubled CO\(_2\) concentration.
The upper row shows these terms at base CO₂ concentration (345 ppm) and the lower row shows their changes in responding to CO₂ doubling (CO₂ × 2 − CO₂ ×1).

At base CO₂ concentration level, we can see that the sum of contributions from the ion drag and the molecular viscosity (Figure 1c) can roughly explain the meridional wind (Figure 1d). The zonal ion drag drives a southward wind in almost the whole thermosphere, with its contribution reaching 80 m s⁻¹. The zonal molecular viscosity drives a northward wind up to 49 m s⁻¹ above ~200 km altitude except for a narrow region below about 250 km near the equator. This result is consistent with Becker and Vadas (2020) which found that the zonal ion drag was the main driver for the zonal mean meridional wind in the thermosphere.

With the doubling of the CO₂ concentration, the southward wind is strengthened by about 20 m s⁻¹ (Figure 1h). This response can be mainly explained by the sum of the responses of the ion drag and molecular viscosity as shown in Figure 1g. Figures 1e and 1f show similar structure and sign as the terms in Figures 1a and 1b (except for northern hemisphere above 250 km), thus revealing that both the ion drag-driven southward wind and the molecular viscosity-driven northward wind strengthened under double CO₂ conditions. For instance, Figure 1e shows southward winds driven by the ion drag strengthens by ~30–40 m s⁻¹, while Figure 1f shows northward wind driven by molecular viscosity strengthens by up to 18 m s⁻¹. Thus, the response in the zonal ion drag is the main driver of the southward wind strengthening at all latitudes.

The zonal ion drag and viscosity cannot, however, explain the peak of the southward wind response between 220 and 280 km altitudes at 40°–60°S. Also, the sum drives northward wind up to 4 m s⁻¹ in 30°–50°S of 160–190 km. To explain these differences, we examined other terms in Equation 7 and find the zonal mean zonal pressure gradient force and meridional advection of zonal wind (the third and fourth terms in Equation 7) have substantial contributions to these differences. Their responses, as shown in Figures 2c and 2d, contribute to the southward wind up to 12 m s⁻¹ in total at latitudes south of ~40°S above 120 km. The vertical advection, geometry term by waves, and artificial term (X) are all below ~2 m s⁻¹ in all regions and hence play a negligible role (not shown).

4.2. Mechanisms for Changes in the Zonal Mean Zonal Wind

As mentioned in the previous section, the ion drag, molecular viscosity, and pressure gradient are dominant forces in the thermosphere, so we investigate these three forces first. Figure 3 shows these terms and their responses to CO₂ doubling, along with their sum and the zonal mean zonal wind.

The sum of the three major terms in Figure 3d roughly represents the zonal wind in Figure 3e, for example, the westward wind (~50 m s⁻¹ at maximum) in the northern hemisphere and the eastward (~30 m s⁻¹ at maximum) at ~48°–45°S. The two features are predominantly attributed to the pressure gradient force (Figure 3a), which drives the westward and eastward winds in the northern and southern hemispheres, respectively, by ~160 m s⁻¹ at maximum. The ion drag (Figure 3b), on the other hand, attenuates the flow driven by the pressure gradient by ~190 m s⁻¹ at maximum. The molecular viscosity (Figure 3c) drives the eastward wind at ~180–250 km and the westward wind above ~250 km by ~50 m s⁻¹ at maximum. It should be noted that magnitudes of the contribution to the zonal wind at ~50°–60°N/S above 130/180 km altitude (regions inside the white lines) have relatively larger uncertainty due to the non-negligible geometric term by the zonal mean wind.

Figure 3j shows the zonal wind response to the doubled CO₂ concentration, with generally eastward wind response up to ~13 m s⁻¹ in the southern hemisphere and weak wind response (less than ~3 m s⁻¹) in the northern hemisphere. Three local structures occur, being the positive peaks poleward of 45°S above 140 km (~13 m s⁻¹) and around 20°S at ~140 km height (~7 m s⁻¹), and the negative peak (~16 m s⁻¹) around 50°N near 110 km. The sum of the three forces in Figure 3i roughly represents the three local peaks, that is, the eastward perturbations above 140 km at latitudes poleward of ~45°S and around 20°S at ~140 km; the westward perturbation around 50°N near 110 km. These peaks are predominantly attributed to the response of the pressure gradient force in Figure 3f, strengthening the eastward/westward winds by 10–120 m s⁻¹ in the southern/northern hemispheres. For the ion drag, its response (Figure 3g) is generally in the opposite direction to that of the pressure gradient force, except for in the southern hemisphere above ~240 km, where it strengthens the pressure-gradient-driven eastward wind by up to ~30 m s⁻¹. The response of the molecular viscosity (Figure 3h) drives eastward winds up to ~110 m s⁻¹ above ~220 km in both hemispheres, but weak eastward wind (up to ~13 m s⁻¹) below. It thus
largely weakens/strengthens the eastward/westward winds driven by the response of the pressure gradient force in the southern/northern hemispheres above 220 km.

Although the response of the sum can partially represent the features of the zonal wind response, there are clear differences above 220 km altitudes in the northern hemisphere and above 160 km at 8°–50°S, where the magnitude of the sum force response in Figure 3i is 10 times larger than the zonal wind change (Figure 3j). These differences are attributed to the meridional advection (Figure 4c) and artificial term, Y (Figure 4d). The responses of the other terms are smaller than the two terms (almost less than 2 m s⁻¹) (not shown). The meridional advection attenuated, up to 8 m s⁻¹, the eastward wind derived by the pressure gradient force in the southern hemisphere. This result indicates that the meridional wave advection has a secondary role on the meridional momentum.

Figure 2. Zonal mean zonal momentum terms divided by zonal mean absolute vorticity under 345 ppm in June (upper row) and their responses to the doubled CO₂ concentration (lower row). (a) and (c) show the pressure gradient force; (b) and (d) show the meridional advection of the zonal wind. The white contours are same as Figure 1.
Figure 3.
balance in the southern (winter) hemisphere. McLandress et al. (2006) also reported the non-negligible meridional wave drag, which is derived from the advection terms (Holton & Hakim, 2013), in the lower thermosphere at lower latitudes. On the other hand, the artificial term, \( Y \), attenuated the zonal wind driven by the pressure gradient force in a height range of 210–270 km in the northern hemisphere (−5 to −15 m s\(^{-1}\)), above ∼250 km in 8°–50°N (−5 to −30 m s\(^{-1}\)), and in 150–260 km in the southern hemisphere (+5 to +23 m s\(^{-1}\)). These large values imply that the zonal wind response has relatively larger ambiguity in these limited regions.

Figure 3. Zonal mean meridional forces divided by the Coriolis parameter under CO\(_2\) x1 in June (a) and (b) and their responses to doubled CO\(_2\) concentration (c) and (d). Positive is eastward. The white contours in the first and third rows indicate a region where the ratio of the geometry term (due to background wind) to the Coriolis force, that is, \( \frac{u^2 \tan \theta}{f} \), is larger than 10% under the base CO\(_2\) concentration. The white contours in the second and fourth rows are the same as in the first and third rows but under either the base CO\(_2\) or doubled CO\(_2\) concentration.

Figure 4. Zonal mean meridional advection of meridional wind and artificial term (\( Y \)) divided by the Coriolis parameter under CO\(_2\) x1 in June (a) and (b) and their responses to doubled CO\(_2\) concentration (c) and (d). Positive is eastward. The white lines are the same as in Figure 3.
5. Mechanisms for Changing the Force Terms

We explored the mechanisms for changes in the zonal mean wind in the thermosphere due to the doubled CO$_2$ concentration in Section 4 by examining the momentum equations. Our analysis showed that the ion drag, molecular viscosity, and meridional pressure gradient changed drastically, with their magnitudes being 3–20 times larger than other terms. The strengthening of the meridional circulation under the doubled CO$_2$ condition is dominantly due to increasing zonal ion drag, while responses in the zonal wind are mainly attributed to changes in the meridional pressure gradient. This section further discusses the physical mechanisms that drive changes in the zonal ion drag and meridional pressure gradient force.

5.1. Strengthening of Zonal Ion Drag Force and Meridional Pressure Gradient Force

The ion drag depends on the ion density and relative velocity between ions and neutral particles (Maeda et al., 1999; Tsuda et al., 2007). The zonal mean zonal ion drag is expressed as

$$\frac{(J \times B)}{\rho} = \frac{\rho_i}{\rho} \nu_{in} (u_i - u).$$

where $\nu_{in}$ is the ion to neutral collision frequency, and $\rho_i$ and $u_i$ are ion density and zonal velocity, respectively. When applying the perturbation expansion, Equation 9 can be rewritten:

$$\frac{\rho_i}{\rho} \nu_{in} (u_i - u) = \left\{ \frac{\nu_{in}}{\rho} \right\} \left\{ \rho_i (\ddot{u} - \ddot{u}) + \rho_i^2 (u_i - u') \right\}. \tag{10}$$

Equation 10 means that the zonal mean ion drag consists of the zonal mean component of the ion density and the relative velocity, and both perturbations. Figure 5c shows responses of the zonal mean ion density. The ion density increases in most regions, with a peak increase at 180–200 km altitude in the southern hemisphere and 220–240 km altitude in the northern hemisphere. Between 20° and 60°S above ∼220 km altitude, the ion density decreased by 50% at maximum. These responses of the ion density are consistent with previous model studies and observations (Danilov & Konstantinova, 2013; Laštovička et al., 2008; H. Liu et al., 2021; Ogawa et al., 2014; Qian et al., 2009; Zhang & Holt, 2011). The strengthened zonal ion drag (Figure 1e) in the northern hemisphere below ∼260 km can be largely attributed to the increase in the ion density (Figure 5c). Above ∼260 km, the zonal ion drag weakens at latitudes between 20° and 40°N, despite of the increase in the ion density by ∼10%. This weakening seems to arise from decreased relative velocity by a few m s$^{-1}$ (Figure 5d). In latitudes equatorward of 20°N, the ion drag also weakens, and this weakening might be attributed to second term in the right-hand side of Equation 10 (which may include longitudinal and local variations of the ion density and the relative velocity). On the other hand, the strengthening of the ion drag in the southern hemisphere seems to be attributed to an increase in the relative velocity, that is, difference between the ion and neutral zonal wind (Figure 5d), except for the altitude range of 160–200 km where the ion drag weakens. This weakening might be also related to the second term in the right-hand side of Equation 10.

The analysis also revealed enhancement of the meridional pressure gradient force, which was dominantly responsible for changes in the zonal mean zonal wind. This strengthening is very likely attributed to a latitudinal asymmetry of the thermosphere density response variation reported by H. Liu et al. (2020), with the density in the southern hemisphere dropping more than that in the northern hemisphere (see figure 1 in H. Liu et al., 2020) due to the downwelling in the southern hemisphere and upwelling in the northern hemisphere related to the strengthened southward meridional circulation in the thermosphere.

5.2. Meridional Advection Caused by Shrinking Effect Versus Dynamical Effect

Results in Section 4 revealed that the meridional advections (fourth term in Equations 7 and 8) also play a role in causing changes in the zonal mean winds, despite of being secondary. Hereby, we would like to examine the relative contributions to the meridional advection from the shrinking effect of the thermosphere due to increasing CO$_2$ concentration and the dynamical effect that is not directly related to thermosphere shrinking.

To distinguish the two effects, we need to compare advections calculated on z-coordinates in Section 4 with those on p-coordinates. This is because advections on z-coordinates respond to changes not only in dynamical...
processes but also to the shrinking of the thermosphere due to increasing CO₂ concentration, while advectons on \( p \)-coordinates reflect only changes caused by the dynamical processes. Figure 6 shows the meridional advection terms and their responses on \( p \)-coordinates. We can see that meridional advection terms (Figures 6c and 6f) above \( 10^{-6} \) hPa (\( \sim 170 \) km altitude) highly resemble those on the \( z \)-coordinates (see Figures 2d and 4c) above \( \sim 170 \) km, suggesting that changes in meridional advection above this altitude are dominantly due to the dynamical effect. On the other hand, differences between both coordinates arise in the pressure range of \( 10^{-4} \) and \( 10^{-6} \) hPa (100–170 km altitudes), suggesting that the shrinking effect is dominant.

We further investigate contributions to the meridional advection from each wave spectrum on the \( z \)-coordinates. The meridional advection of zonal and meridional wind on the \( z \)-coordinates can be written as follows using Fourier transformation:

\[
\frac{1}{r} \frac{\partial u'}{\partial \theta} = \sum_{m=1}^{N} \sum_{n=m}^{\Omega} v_{m,n} \frac{1}{r} \frac{\partial u_{m,n}'}{\partial \theta},
\]

(11)

\[
\frac{1}{r} \frac{\partial v'}{\partial \theta} = \sum_{m=1}^{N} \sum_{n=m}^{\Omega} v_{m,n} \frac{1}{r} \frac{\partial v_{m,n}'}{\partial \theta}.
\]

(12)
where \( n \) and \( \omega \) are a zonal wavenumber and frequency. \( N \) and \( \Omega \) are the maximum wavenumber and frequency.

Our analysis reveals that the meridional advection terms and their changes are mainly attributed to changes in the diurnal tidal component with zonal wavenumber 1 (DW1). As shown in Figure 7, the advection caused by DW1 amounts to \( \sim 70\% \) of the total advection shown in Figures 2 and 4. For instance, the advection of zonal wind by DW1 in Figure 7c drives a southward wind difference of \( \sim 4 \) m s\(^{-1}\) above \( \sim 130 \) km poleward of 30°S, in comparison to the total wind change of \( \sim 6 \) m s\(^{-1}\) in Figure 2d. The advection of meridional wind by DW1 in Figure 7d drives a southward wind change of \( \sim 6 \) m s\(^{-1}\) at maximum, in comparison to the total wind change of \( \sim 9 \) m s\(^{-1}\) in Figure 4c. This enhanced DW1 advection is consistent with an enhanced amplitude of DW1 reported by H. Liu et al. (2020). This DW1 is probably the evanescent diurnal tide excited by in situ absorption of solar extreme ultraviolet and ultraviolet radiation, because DW1 excited in the troposphere cannot penetrate above \( \sim 120 \) km (Forbes & Garrett, 1979; Richmond, 1979). One of the possible mechanisms for the enhanced DW1 is an interaction between DW1 and the zonal wind (Lieberman et al., 2013), but we need further study to verify this.

6. Discussion and Conclusion

We explored the mechanisms for responses of thermospheric zonal mean winds to a doubling of the CO\(_2\) concentration around June solstice by examining the momentum balance of different terms in the momentum equations (see Equations 7 and 8). The analysis reveals that the ion drag, molecular viscosity, and meridional pressure gradient forces are the primary terms that determine responses in the winds, with the meridional advection of tidal winds playing a secondary role. The results demonstrate that changes in large-scale zonal mean winds in the thermosphere are caused by the following sequential processes: (a) The CO\(_2\) increase causes enhancement of ion density in almost the whole northern hemisphere and below \( \sim 210 \) km in the southern hemisphere. (b) This enhancement strengthens the zonal ion drag in the northern hemisphere, resulting in a faster meridional circulation. (c) This faster circulation causes the latitudinal asymmetry of the density which increases the meridional pressure gradient. (d) This increased pressure gradient strengthens the zonal mean zonal wind in the southern...
hemisphere; this consequently strengthens the zonal relative velocity between ions and neutrals and the zonal ion drag. Therefore, we conclude that the increased ion density in the northern hemisphere is the underlying trigger for the wind responses, that is, the ionosphere responses to the CO₂ cooling induce variations in the thermosphere circulation. This wind response might feedback onto the ionosphere, but further studies are needed. This study demonstrates the importance of ionosphere-thermosphere interaction via ion drag in the collective responses to lower atmosphere forcing, making the interaction a highly necessary process in whole atmosphere models. Although the present study investigated the wind response in June, the response in December could be attributed to the sequential processes because the pressure gradient and meridional circulation are strong as in June. On the other hand, a different process possibly works to the wind during an equinox, because the pressure gradient and meridional circulation should be weak. Indeed, H. Liu et al. (2020) reported a different wind response in September from that in June.

Finally, we would like to note two ambiguities in our study. First, the response of the term $\gamma$ in Equation 8 is up to $\sim30 \text{ m s}^{-1}$, causing some ambiguity in the zonal wind response. It would be useful to compare our GAIA simulation with other model simulations (e.g., The Thermosphere-Ionosphere-Mesosphere Electrodynamics-General Circulation Model, i.e., TIME-GCM; The Whole Atmosphere Community Climate Model with thermosphere and ionosphere extension, i.e., WACCM-X). Second, recent studies (Becker & Vadas, 2018, 2020; Vadas &
Becker, 2019) show that secondary or high-order GWs caused by local body forces transport non-neglectable momentum into the thermosphere. Since the current GAIA model does not capture such GWs, it will lead to some ambiguity in the advection terms. Further work is needed to understand their role in response to increased CO₂ concentration.

Data Availability Statement

Our GAIA simulation data can be available at https://doi.org/10.5281/zenodo.5944788.

References

Andrews, D. G., Holton, J. R., & Leovy, C. B. (1987). Middle atmosphere dynamics (Vol. 40). Academic Press.

Banks, P. M., & Kockarts, G. (1973). Aeronomy, Part B (p. 355). New York: Elsevier.

Becker, E., & Vadas, S. L. (2018). Secondary gravity waves in the winter mesosphere: Results from a high-resolution global circulation model. Journal of Geophysical Research: Atmospheres, 123, 2605–2627. https://doi.org/10.1002/2017JD027460

Becker, E., & Vadas, S. L. (2020). Explicit global simulation of gravity waves in the thermosphere. Journal of Geophysical Research: Space Physics, 125, e2020A028034. https://doi.org/10.1029/2020JA028034

Brum, C. G. M., Tepley, C. A., Fentzke, J. T., Robles, E., dos Santos, P. T., & Gonzalez, S. A. (2012). Long-term changes in the thermospheric neutral winds over Arecibo: Climatology based on over three decades of Fabry-Perot observations. Journal of Geophysical Research, 117, A00H14. https://doi.org/10.1029/2011JA016458

Chapman, S., & Lindzen, R. S. (1970). Atmospheric tides. New York: Gordon and Breach.

Cnossen, I. (2020). Analysis and attribution of climate change in the upper atmosphere from 1950 to 2015 simulated by WACCM-X. Journal of Geophysical Research: Space Physics, 125, e2020A026623. https://doi.org/10.1029/2020JA026623

Danilov, A. D., & Konstantinova, A. V. (2013). Trends in the F2 layer parameters at the end of the 1900s and the beginning of the 2000s. Journal of Geophysical Research: Atmospheres, 118, 5947–5964. https://doi.org/10.1002/jgrd.50501

Emmert, J. T., Fejer, B. G., Shepherd, G. G., & Solheim, B. H. (2004). Average nighttime F region disturbance neutral winds measured by UARS WINDII: Initial results. Geophysical Research Letters, 31, L22807. https://doi.org/10.1029/2004GL021611

Emmert, J. T., Picone, J. M., Lean, J. L., & Knowles, S. H. (2004). Global change in the thermosphere: Compelling evidence of a secular decrease in density. Journal of Geophysical Research, 109, A02301. https://doi.org/10.1029/2003JA010176

Forbes, J. M. (2007). Dynamics of the thermosphere. Journal of the Meteorological Society of Japan Series II, 85, 193–213.

Forbes, J. M., & Garrett, H. B. (1979). Theoretical studies of atmospheric tides. Reviews of Geophysics, 17(8), 1951–1981. https://doi.org/10.1029/RG017i008p01951

Holton, J. R., & Hakim, G. J. (2013). An introduction to dynamic meteorology (5th ed.). Waltham, MA: Academic Press. https://doi.org/10.1016/C2009-0-6394-8

Jin, H., Miyoshi, Y., Fujiwara, H., Shinagawa, H., Terada, K., Terada, N., et al. (2011). Vertical connection from the tropospheric activities to the ionospheric longitudinal structure simulated by a new Earth’s whole atmosphere-ionosphere coupled model. Journal of Geophysical Research, 116, A01316. https://doi.org/10.1029/2010JA015925

Keating, G. M., Tolson, R. H., & Bradford, M. S. (2000). Evidence of long term global decline in the Earth's thermospheric densities apparently related to anthropogenic effects. Geophysical Research Letters, 27, 1523–1526. https://doi.org/10.1029/2000GL007371

Laštovička, J. (2021). Long-term trends in the upper atmosphere. In W. Wang, Y. Zhang, & L. J. Paxton (Eds.), Upper atmosphere dynamics and energetics. https://doi.org/10.1002/9781119185631.ch17

Laštovička, J., Akmaev, R. A., Beig, G., Bremer, J., & Emmert, J. T. (2006). Global change in the upper atmosphere. Science, 314, 1253–1254. https://doi.org/10.1126/science.1135134

Laštovička, J., Akmaev, R. A., Beig, G., Bremer, J., Emmert, J. T., Jacoby, C., et al. (2008). Emerging pattern of global change in the upper atmosphere and ionosphere. Annals of Geophysics, 26, 1255–1268. https://doi.org/10.5194/angeo-26-1255-2008

Lieberman, R. S., Akmaev, R. A., Fuller–Rowell, T. J., & Doornbos, E. (2013). Thermospheric zonal mean winds and tides revealed by CHAMP Geophysical Research Letters, 40, 2439–2443. https://doi.org/10.1002/2013GL054081

Lindzen, R. S. (1981). Turbulence and stress owing to gravity wave and tidal breakdown. Journal of Geophysical Research, 86(C10), 9707–9714. https://doi.org/10.1029/JC086iC10p09707

Liu, H.-L., Foster, B. T., Hagan, M. E., McNerney, J. M., Maute, A., Qian, L., et al. (2020). Southern winter stratosphere: Assessment and attribution. Journal of Geophysical Research: Space Physics, 125, e2020A028612. https://doi.org/10.1029/2020JA028612

Liu, H., Tao, C., Jin, H., & Abe, T. (2021). Geomagnetic activity effects on CO₂-driven trend in the thermosphere and ionosphere: Ideal model experiments with GAIA. Journal of Geophysical Research: Space Physics, 126(1), e2020A028741. https://doi.org/10.1029/2020JA028741

Liu, H.-L., Foster, B. T., Hagan, M. E., McNerney, J. M., Maute, A., Qian, L., et al. (2010). Thermosphere extension of the whole atmosphere community climate model. Journal of Geophysical Research, 115, A12302. https://doi.org/10.1029/2010JA015586

Maeda, S., Fujiwara, H., & Nozawa, S. (1999). Momentum balance of dayside F region neutral winds during geomagnetically quiet summer days. Journal of Geophysical Research, 104(A9), 19871–19879. https://doi.org/10.1029/1999JA900224

Marco, F. A., Wise, J. O., Kendra, M. J., Grossbard, N. J., & Bowman, B. R. (2005). Detection of a long-term decrease in thermospheric neutral density. Geophysical Research Letters, 32, L04103. https://doi.org/10.1029/2004GL021269

McFarlane, N. A. (1987). The effect of orographically excited gravity wave drag on the general circulation of the lower stratosphere and troposphere. Journal of the Atmospheric Sciences, 44(14), 1775–1800. https://doi.org/10.1175/1520-0469(1987)044<1775:TOEGWJ>2.0.CO;2

McLundress, C., Ward, E. V., Fomichev, V. I., Semenik, K., Beagley, S. R., McFarlane, N. A., & Shepherd, T. G. (2006). Large-scale dynamics of the mesosphere and lower thermosphere: An analysis using the extended Canadian Middle Atmosphere Model. Journal of Geophysical Research, 111, D17111. https://doi.org/10.1029/2005JD006776

Miyahara, S., Yoshida, Y., & Miyoshi, Y. (1993). Dynamic coupling between the lower and upper atmosphere by tides and gravity waves. Journal of Geophysical Research: Atmospheres, 98(D2), 3581–3598. https://doi.org/10.1029/92JD01771

Miyoshi, Y., Fujiwara, H., Jin, H., & Shinagawa, H. (2015). Impacts of sudden stratospheric warming on general circulation of the thermosphere. Journal of Geophysical Research: Space Physics, 120, 10897–10912. https://doi.org/10.1002/2015JA021894

Acknowledgments

H. L. acknowledges support by JSPS KAKENHI Grants 18H01270, 17K0095, 20H00197, and JRP-LEAD with DFG program (IPSSIRP 2018/1602). C. T. is supported by JSPS KAKENHI 19K03942. M. K. is supported by JSPS KAKENHI 19K023465, 22H00331, 2022 Research Start Program 202204, and the SCAR Fellowship Programme.
Moritz, H. (2000). Geodetic reference system 1980. *Journal of Geodesy, 74*, 128–133. https://doi.org/10.1007/s001900050278

Nappo, C. J. (2012). *An introduction to atmospheric gravity waves* (2nd ed.). New York, NY: Academic Press.

Ogawa, Y., Motoba, T., Buchert, S. C., Hågström, I., & Nozawa, S. (2014). Upper atmosphere cooling over the past 33 years. *Geophysical Research Letters, 41*, 5629–5635. https://doi.org/10.1002/2014GL060591

Qian, L., Burns, A. G., Solomon, S. C., & Roble, R. G. (2009). The effect of carbon dioxide cooling on trends in the F2-layer ionosphere. *Journal of Atmospheric and Solar-Terrestrial Physics, 71*(14–15), 1592–1601. https://doi.org/10.1016/j.jastp.2009.03.006

Qian, L., Laitovička, J., Roble, R. G., & Solomon, S. C. (2011). Progress in observations and simulations of global change in the upper atmosphere. *Journal of Geophysical Research, 116*, A00H03. https://doi.org/10.1029/2010JA016317

Richmond, A. D. (1979). Ionospheric wind dynamo theory: A review. *Journal of Geomagnetism and Geoelectricity, 31*(3), 287–310.

Rishbeth, H., & Roble, R. G. (1992). Cooling of the upper atmosphere by enhanced greenhouse gases—Modelling of thermospheric and ionospheric effects. *Planetary and Space Science, 40*(7), 1011–1026. https://doi.org/10.1016/0032-0633(92)90141-A

Roble, R. G., & Dickinson, R. E. (1989). How will changes in carbon dioxide and methane modify the mean structure of the mesosphere and thermosphere? *Geophysical Research Letters, 16*(12), 1441–1444. https://doi.org/10.1029/GL016i012p01441

Sato, K., Yasui, R., & Miyoshi, Y. (2018). The momentum budget in the stratosphere, mesosphere, and lower thermosphere. Part I: Contributions of different wave types and in situ generation of Rossby waves. *Journal of the Atmospheric Sciences, 75*(10), 3613–3633. https://doi.org/10.1175/jas-d-17-0336.1

Saunders, A., Lewis, H., & Swinerd, G. (2011). Further evidence of long-term thermospheric density change using a new method of satellite ballistic coefficient estimation. *Journal of Geophysical Research, 116*, A00H10. https://doi.org/10.1029/2010JA016358

Solomon, S. C., Liu, H.-L., Marsh, D. R., McInerney, J. M., Qian, L., & Viti, F. M. (2018). Whole atmosphere simulation of anthropogenic climate change. *Geophysical Research Letters, 45*, 1567–1576. https://doi.org/10.1029/2017GL076950

Solomon, S. C., Qian, L., & Roble, R. G. (2015). New 3-D simulations of climate change in the thermosphere. *Journal of Geophysical Research: Space Physics, 120*, 2183–2193. https://doi.org/10.1002/2014JA020886

Tepley, C. A., Robles, E., Garcia, R., Santos, P. T., Brum, C. M., & Burnside, R. G. (2011). Directional trends in thermospheric neutral winds observed at Arecibo during the past three solar cycles. *Journal of Geophysical Research, 116*, A00H06. https://doi.org/10.1029/2010JA016172

Tsuda, T. T., Nozawa, S., Brekke, A., Ogawa, Y., Motoba, T., Roble, R., & Fuji, R. (2007). An ion drag contribution to the lower thermospheric wind in the summer polar region. *Journal of Geophysical Research, 112*, A06319. https://doi.org/10.1029/2006JA011785

Vadas, S. L. (2007). Horizontal and vertical propagation and dissipation of gravity waves in the thermosphere from lower atmospheric and thermospheric sources. *Journal of Geophysical Research, 112*, A06305. https://doi.org/10.1029/2006JA011845

Vadas, S. L., & Becker, E. (2019). Numerical modeling of the generation of tertiary gravity waves in the mesosphere and thermosphere during strong mountain wave events over the Southern Andes. *Journal of Geophysical Research: Space Physics, 124*, 7687–7718. https://doi.org/10.1029/2019JA026694

Vadas, S. L., & Fritts, D. C. (2004). Thermospheric responses to gravity waves arising from mesoscale convective complexes. *Journal of Atmospheric and Solar-Terrestrial Physics, 66*(6–9), 781–804. https://doi.org/10.1016/j.jastp.2004.01.025

Zhang, S.-R., & Holt, J. M. (2011). Millstone Hill ISR observations of upper atmospheric long-term changes: Height dependency. *Journal of Geophysical Research, 116*, A00H05. https://doi.org/10.1029/2010JA016414