Impacts of Increasing CO₂ on Diurnal Migrating Tide in the Equatorial Lower Thermosphere

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Abstract. We investigate impacts of increased CO_2 concentration on migrate diurnal tide (DW1). A future climate
simulation is conducted using a WACCM-X model, with surface CO_2 levels increasing according to the RCP 8.5
10   scenario. The DW1 (1,1) mode, a propagating tide peaking near the equator, exhibits a positive trend of ~+1% per
decade in a range of 20–70 km, and a negative trend of ~-2% per decade in a range of 90–110 km. The positive
trend is likely driven by depression in atmospheric density in the mesosphere and enhanced equatorial convective
activity, while the negative trend appears to result from increased eddy diffusion in the mesosphere, which
overwhelms the positive trend. Two potential mechanisms may explain the negative trend. First, increasing CO_2
15   enhances mesospheric stability, reducing tidal vertical wavelengths. In our simulation, equatorial temperatures
around ~50-70 km become cooler than those in ~70-90 km. This strong cooling could be linked to CO_2 mixing and
transport, as well as the contraction of the mesospheric ozone layer due to CO_2-induced cooling. Second, stronger
convective activity intensifies gravity wave generation, increasing gravity wave diffusion in the mesosphere. This
strong convective activity also likely intensifies the tide below ~70 km. While our positive DW1 trend is consistent
20   with McLandress and Fomichev (2006), the negative trend in the lower thermosphere contrasts with their results.
This discrepancy might arise because their model used a time-independent diffusion coefficient, whereas WACCM-
X accounts for CO_2-driven changes in gravity wave diffusion.

1.   Introduction

25   Anthropogenic CO_2 emissions from fossil fuel combustion and industrial processes have risen steadily
since the Industrial Revolution and are projected to continue increasing in the future (IPCC, 2023). This rise has
significantly elevated atmospheric CO_2 concentrations, contributing to global warming in the troposphere. Global
warming has altered geographical precipitation patterns, including their frequency and intensity, as well as
atmospheric circulation patterns (e.g., Arias et al., 2021; Chou & Neelin, 2004; Feng et al., 2019; Lau et al., 2013).
30   On the other hand, atmospheric layers above the tropopause, such as the stratosphere, mesosphere, thermosphere,
and ionosphere, have been cooling as CO_2 concentration has increased, and their altitudes have descended (e.g.,
Akmaev and Fomichev, 1998; Arias et al., 2021; Cnossen, 2020; Emmert et al., 2010; Emmert, Fejer, et al., 2004;
Emmert, Picone, et al., 2004; Garcia, 2021; Garcia et al., 2019; Jonsson et al., 2004; Keating et al., 2000; Kogure et
al., 2022; Laštovička, 2021; Laštovička et al., 2008, 2012; H. Liu et al., 2020, 2021; Marcos et al., 2005; Ogawa et
35   al., 2014; Qian et al., 2011; Ramesh & Sridharan, 2018; Roble & Dickinson, 1989; Zhang et al., 2011). Akmaev and
Fomichev (1998) demonstrated that cooling/contraction of the atmosphere can lead to apparent warming at specific
altitudes in the lower thermosphere (~100–120 km) due to the large temperature gradient therein. Similarly, the
ionosphere also descends in response to CO_2 cooling, altering the vertical profiles of ion and electron density (see
Figure 16.5 in Laštovička, 2021).
40   In addition to changes in temperature, the dynamical effects of increasing CO_2 on the thermosphere have
been demonstrated by Liu et al. (2020), showing for the first time an enhanced circulation and significant changes in
thermospheric tidal activities. Since thermal tides play a key role in thermosphere–ionosphere dynamics,
understanding the impact of rising CO_2 levels on these tides is essential for predicting thermospheric climate change.
Tides are classified into two categories based on their dynamical characteristics: trapped and propagating tides
45   (Chapman & Lindzen, 1970). Trapped tides from in-situ forcing dominate in the middle and upper thermosphere
(above ~140 km altitude), whereas tides in the lower thermosphere primarily originate in the troposphere and
stratosphere and propagate upward (Yamazaki et al., 2024; Yamazaki and Siddiqui, 2014). Using the whole

atmosphere model Ground-to-topside Atmosphere Ionosphere model for Aeronomy (GAIA), Liu et al. (2020) found an enhancement of the migrating tides (DW1) below 200 km, and a reduction of the migrating semidiurnal tides (SW2) throughout most of the thermosphere. Ma et al. (2025) confirmed similar tidal trends in a long-term future projection simulation of WACCM-X (The Whole Atmosphere Community Climate Model with thermosphere and ionosphere extension). Regarding the lower thermosphere, McLandress and Fomichev (2006) examined the impacts of doubled CO_2 on propagating DW1 tides, which primarily occur in the equatorial MLT (mesosphere–lower thermosphere) region. Using a linear tidal model, they compared three scenarios: (1) present-day CO_2 levels, (2) doubled CO_2 with present-day sea surface temperatures, and (3) doubled CO_2 with sea surface temperatures adjusted accordingly. These scenarios were based on present-day observations and simulations conducted with the Canadian Middle Atmosphere Model (CMAM). Their results suggested that doubling CO_2 increases tropospheric water vapor and its absorption of solar heating, amplifying DW1 tides in the MLT by ~1 K (10–15%). However, their linear tidal model did not incorporate changes in gravity wave (GW) diffusion, despite its significant role in tidal dissipation, because it used a time-independent eddy diffusion coefficient, which is independent of GW drag (McLandress, 2002). Additionally, since propagating DW1 tides are generated by tropospheric disturbances and modulated by stratospheric and mesospheric background conditions, uncertainties in lower atmospheric states directly affect predictions of future DW1 variability.

This study investigates the impact of increasing CO_2 on propagating DW1 tides using the state-of-the-art model, WACCM-X. We used a 2° horizontal resolution version of WACCM-X. Although this model cannot resolve most gravity waves, it includes both orographic and non-orographic gravity wave parameterizations.

Section 2 describes the specifications of WACCM-X, gravity wave parameterizations, simulation setup, and our analysis method, specifically the Hough mode decomposition. Section 3 presents the results, showing a positive trend in propagating DW1 tides in the stratosphere and a negative trend in the lower thermosphere. Section 4 discusses potential mechanisms for enhanced tidal dissipation in the MLT region. Finally, Section 5 summarizes our findings, compares them with those of McLandress and Fomichev (2006), and discusses the limitations of this study. Unless otherwise stated, "tides" hereafter refers to propagating DW1 tides.

2. Data and Analysis

2.1. WACCM-X Simulation

This study uses the same long-term simulation as those used in Ma et al. (2025) and Pedatella et al. (2025). Briefly, it is a 90-year simulation (2000–2089) using CESM/WACCM-X version 2.2 with coupled ocean. The model has latitudinal and longitudinal resolutions of 1.9° × 2.5°, respectively. Its vertical resolution decreases with altitude, transitioning from 0.16 density scale height at 100 hPa to 0.25 density scale height at 1 hPa. The model time step is 15 minutes, and we used monthly mean output parameters for our analysis. WACCM-X simulates the entire atmosphere, from Earth's surface to the thermosphere and ionosphere, with a model top at ~500-700 km altitude. It includes orographic and non-orographic GW parameterizations, accounting for three GW sources: deep convection (Alexander et al., 2021; Beres et al., 2005), frontogenesis (Richter et al., 2010), and winds over complex terrain (Garcia et al., 2017; Scinocca and McFarlane, 2000). WACCM-X also includes a cumulus convection parameterization (Zhang and McFarlane, 1995), which is coupled with the parameterization for convectively-generated GWs (Beres et al., 2005). The model also solves chemical reactions for five ion species, electrons, and 74 neutral species, including ozone. A detailed description of the dynamical processes in WACCM-X (version 2) is provided by H. L. Liu et al. (2018).

90   In this simulation, the horizontally uniform CO_2 concentration is specified near the surface and follows historical observations up to 2014, after which it follows the Representative Concentration Pathway 8.5 (RCP 8.5) scenario. Above the surface, CO_2 concentrations vary due to atmospheric transportation and eddy diffusion. Although the simulation runs until the end of 2089, we analyze only the data before July 2069, as the surface CO_2 concentration exceeds the upper limit of the Fomichev non-LTE CO_2 cooling scheme (720 ppm). Figure 1a shows the global mean CO_2 concentrations, smoothed over one-year, at the surface and altitudes of 50 km and 100 km,

95   respectively. Solar cycle activity follows the Coupled Model Intercomparison Project (CMIP6) (O'Neill et al., 2016); that is, it is specified using historical observations up to 2014, and from 2015 onward, it is simulated by rewinding the solar forcing to the 1850 levels (Figure 1b).
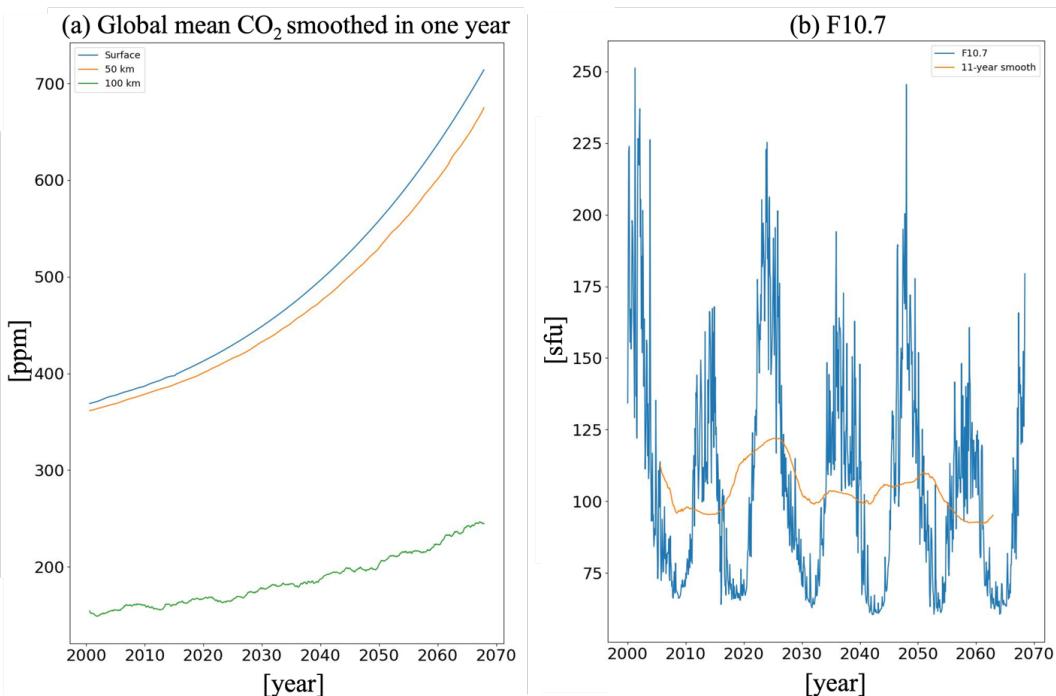
Figure 1. (a) Global mean CO_2 concentrations, smoothed over one year. The blue, orange, and green lines represent CO_2 concentrations at the surface, 50 km, and 100 km altitudes, respectively. (b) F10.7 solar flux (blue) and its 11-

100   year smoothed trend (orange).

2.2. DW1 (1, 1) Mode

     Monthly mean temperature perturbations of the DW1 tide were derived from the output monthly mean diurnal components (the parameter names are "Temperature 24hr. cos and sin coeff.") using a discrete Fourier

105   transform. These perturbations were then convolved with the (1,1) Hough mode function to derive the (1,1) mode amplitudes on pressure coordinates. These amplitudes were subsequently interpolated to geometric height for each month. Unless otherwise stated, other parameters are also processed in the same manner. The Hough mode functions are solutions of Laplace's tidal equation under the assumptions of an isothermal atmosphere with no background wind. The (1,1) Hough mode is predominantly generated by solar heating of tropospheric water vapor and

110   propagates upward into the lower thermosphere. This mode is dominant in the equatorial region, accounting for over ~90% of DW1 amplitudes, as its energy is concentrated within 30°N/S, with peak temperature amplitudes at the equator. The Hough function was calculated using the normalized Associated Legendre Polynomial (ALP) expansion method (Groves, 1981), implemented through a Python module developed in this study. This module is based on the MATLAB program developed by Wang et al. (2016), but it also includes functionality for calculating

115   the Hough modes of D0 tidal waves, which cannot be computed in the original MATLAB program.

3. Results

Figure 2a shows the amplitudes of the (1,1) Hough mode in the 70-110 km geometric height range, representing the MLT region. The output geometric height (z^*) is defined in the WACCM-X simulation as:

$$z^* = \frac{R_e}{R_e - z_{GP}} \tag{1}$$

where R_e and z_{GP} are the Earth's radius (6371 km) and geopotential height, respectively (Neale et al., 2010). The amplitudes were smoothed using a 2-year boxcar window because the stratospheric quasi-biennial oscillations (QBO) significantly modulate tidal wave amplitudes (Hagan et al., 1999; Liu et al, 2017, Kogure and Liu, 2021; Xu et al., 2009). It should be noted that the QBO is specified by climatology. The amplitudes peak between 100 and 110 km, and these peak altitudes remain constant until 2069, while the peak amplitudes gradually decrease over time. For example, maximum amplitudes were 16–18 K during 2001–2010 but decreased to 14–16 K during 2058–2067, representing a reduction of approximately 2 K(~10%). This negative trend can also be seen on pressure coordinates (Figure 2b). Because of the facilitation of comparison with observations, this study focuses on the results of geometric height coordinates. Figures 2c–g present time series of the monthly mean tidal amplitudes averaged over the 20–30 km, 30–50 km, 50–70 km, 70–90 km, and 90–110 km altitude ranges (green crosses). The 2-year smoothed amplitudes are shown in black, while their linear trends are depicted in red. Table 1 summarizes the slopes and their standard errors for the linear fits. Here, the standard error denotes the one-sigma (~68%) confidence interval of the regression, which is calculated by SciPy module (Virtanen et al., 2020). Notably, the amplitudes in the 90–110 km range significantly decrease over time, while those below 70 km significantly increase. In the 70 and 90 km range, the magnitude of the slope is 4–5 times smaller than its standard error, indicating no significant trend.

This result suggests that tidal source activity in the troposphere (solar heating due to water vapor absorption and latent heating) intensifies, leading to strong DW1 tides below ~70 km. Conversely, above ~70 km, tidal dissipation increases, offsetting the positive effects of enhanced source activity and reducing the amplitude above ~ 90 km.
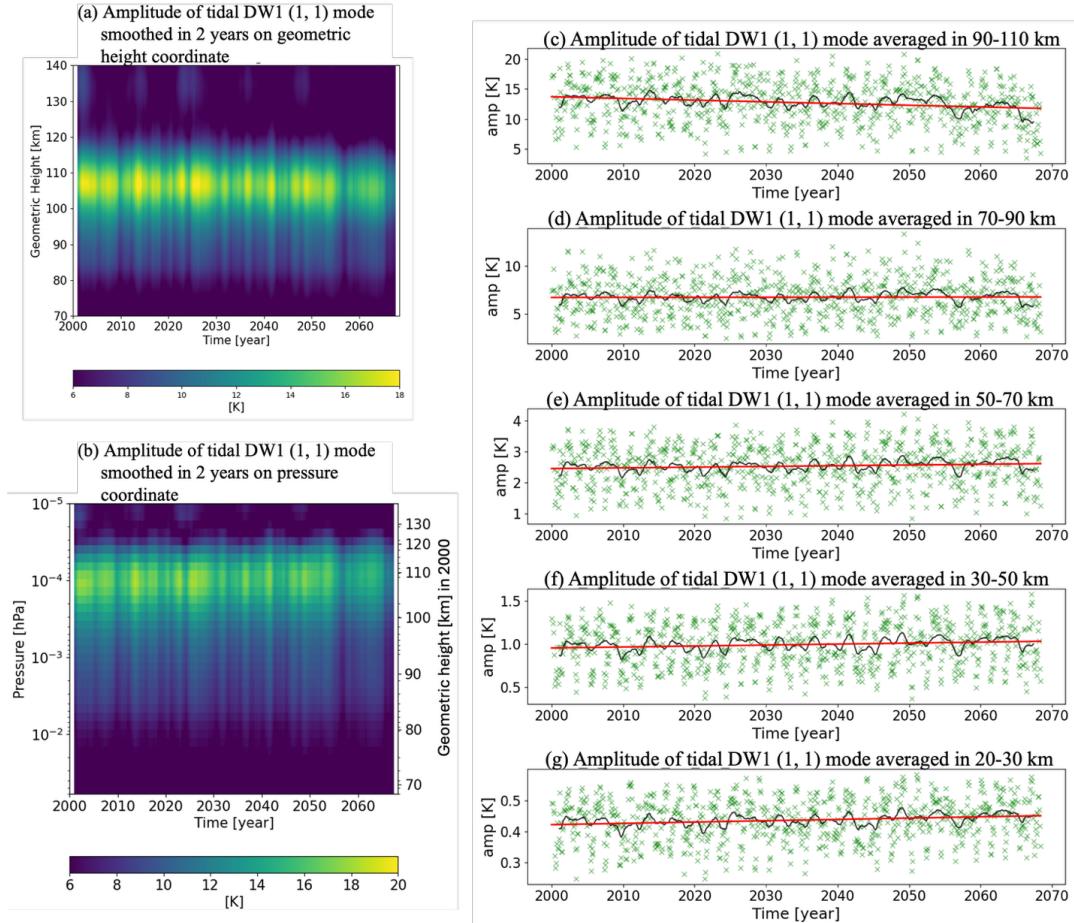
Figure 2. (a) Amplitudes of the (1,1) Hough mode smoothed over two years on geometric height (a) and pressure coordinates (b), respectively. The right-axis in (b) shows the geometric height in 2000 corresponding to the left-axis (pressure) for reference. Note that geometric height sinks with atmospheric cooling. (c–g) Time series of monthly mean amplitudes averaged over 20–30 km, 30–50 km, 50–70 km, 70–90 km, and 90–110 km. The green crosses represent monthly values; the black lines denote 2-year smoothed values; the red lines represent linear trends.

Height [km]	Trend of the amplitude [K/year]	Standard error of trend [K/year]
20-30	4.2×10^{-4}	1.2×10^{-4}
30-50	1.1×10^{-3}	4.6×10^{-4}
50-70	2.3×10^{-3}	1.2×10^{-4}
70-90	8.0×10^{-4}	3.7×10^{-3}
90-110	-2.8×10^{-2}	6.3×10^{-3}

Table 1. Slopes and their standard errors of the linear fit shown in Figure 2(c-g).

To further investigate the impact of increasing CO_2 on the tidal vertical propagation, particularly above ~70 km, we compared the tidal amplitudes during two periods: January 2003–December 2013 and December 2050–

150   November 2061. These periods were selected because they have similar mean F10.7 values and standard deviations (95.9 ± 26.6 sfu in 2003–2013 vs. 95.8 ± 23.4 sfu in 2050–2061). However, the mean CO_2 concentration at the surface in 2050–2061 is 608 ppm, ~58% higher than in 2003–2013. Figure 3a illustrates the ratio of the amplitudes between 2003–2013 and 2050-2061. Across most of the 40-82 km range, amplitude in 2050-2061 are significantly larger (by up to ~9%) than those in 2003-2013. However, above ~72 km, the ratio gradually decreases with altitude

155   and drops below 100% at ~82 km, indicating that amplitudes in 2050-2061 are smaller than those in 2003-2013 above this altitude.

       The ratio shown in Figure 3a depends on variations in tidal source activity in the troposphere, atmospheric density, and propagation conditions. The vertical structure of wave amplitude, $A_{(z)}$, can be approximately described based on Forbes and Vicent (1989):

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$$A_{(z)} = A_{(z_0)} \sqrt{\frac{\rho_{(z_0)}}{\rho_{(z)}}} \exp \int_{z_0}^{z} -\sigma_{i_{(z)}} \, dz \,. \tag{2}$$

Here, z and z_0 represent altitude and the altitude of the wave source (i.e., the troposphere for the DW1 tide), respectively. $\rho_{(z)}$ denotes atmospheric density, and $\sigma_{i_{(z)}}$ is the damping factor due to dissipation. Using equation (1), the ratio of tidal amplitude at z between 2003–2013 and 2050–2061 can be expressed as:

$$\frac{A_{2050-2061_{(z)}}}{A_{2003-2013_{(z)}}} = \frac{A_{2050-2061_{(z_0)}}}{A_{2003-2013_{(z_0)}}} \sqrt{\frac{\rho_{2050-2061_{(z_0)}}}{\rho_{2003-2013_{(z_0)}}}} \sqrt{\frac{\rho_{2003-2013_{(z)}}}{\rho_{2050-2061_{(z)}}}} \exp \int_{z_0}^{z} -(\sigma_{i_{2050-2061_{(z)}}} + \sigma_{i_{2003-2013_{(z)}}}) \, dz \,. \tag{3}$$

165   In the WACCM-X simulation, the average value of $\sqrt{\frac{\rho_{2050-2061_{(z_0)}}}{\rho_{2003-2013_{(z_0)}}}}$ in 0–17 km is $100.2 \pm 0.6\%$, which is almost unity. Thus, equation (2) simplifies to:

$$\frac{A_{2050-2061_{(z)}}}{A_{2003-2013_{(z)}}} = \frac{A_{2050-2061_{(z_0)}}}{A_{2003-2013_{(z_0)}}} \sqrt{\frac{\rho_{2003-2013_{(z)}}}{\rho_{2050-2061_{(z)}}}} \exp \int_{z_0}^{z} -(\sigma_{i_{2050-2061_{(z)}}} + \sigma_{i_{2003-2013_{(z)}}}) \, dz \,. \tag{4}$$

Therefore, the amplitude ratio shown in Figure 3a can be explained by changes in the tidal amplitude at z_0 (source activity), the atmospheric density, and tidal damping factor due to dissipation (propagation condition). Figure 3b

170   shows the square root of the atmospheric density ratio between 2003–2013 and 2050–2061, $\sqrt{\frac{\rho_{2003-2013_{(z)}}}{\rho_{2050-2061_{(z)}}}}$. This ratio gradually increases with altitude above 50 km, reaching by ~4% at ~72 km. A similar increasing trend is observed in the amplitude ratio shown in Figure3a. These results suggests that the tidal positive trend in 50–70 km is driven not only by enhanced source activity but also by the decrease in atmospheric density in the mesosphere. However, above ~72 km, the amplitude ratio decreases with altitude, even though $\frac{\rho_{2003-2013_{(z)}}}{\rho_{2050-2061_{(z)}}}$ continues to rise. These vertical

175   variations suggest that tidal dumping factor ($\sigma_{i_{2050-2061_{(z)}}}$) increases significantly above ~72 km in 2050-2061.

This enhanced damping counteracts the tidal amplification caused by the increased source activity and reduced density, leading to a decline in tidal amplitude near ~82 km. As a result, a negative tidal trend is projected in the MLT region.
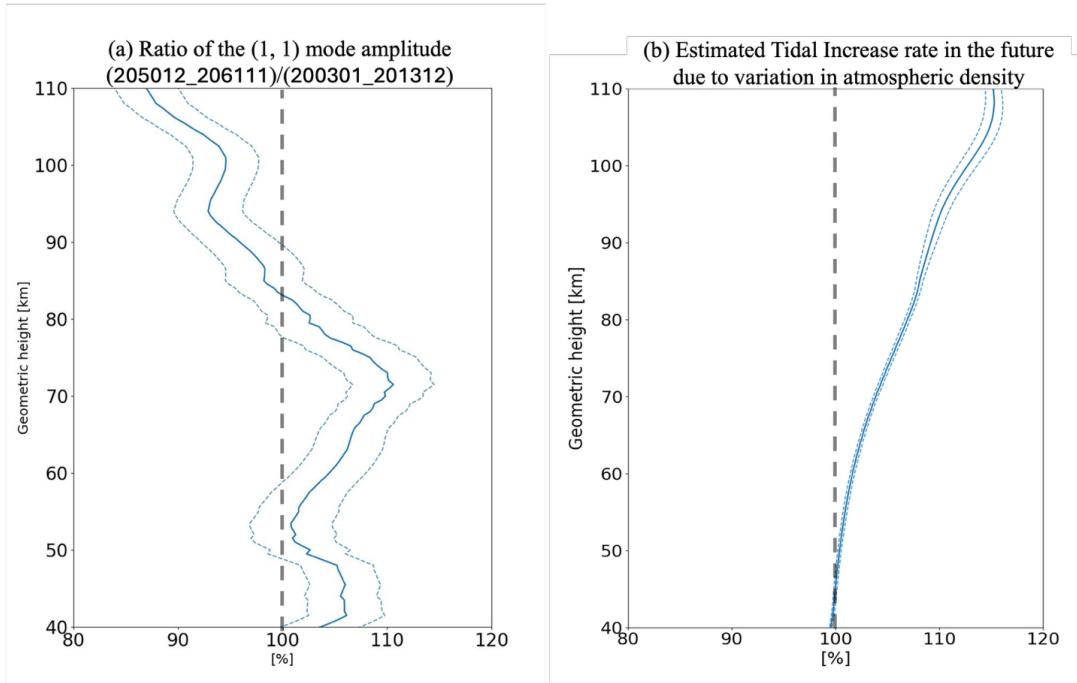
Figure 3. (a) Percentage ratio of the tidal amplitude, averaged over 2050-2061 and normalized by the average in
2003–2013. (b) Estimated future tidal increase rate due to changes in atmospheric density ($\frac{\rho_{2003-2013(z)}}{\rho_{2050-2061(z)}}$). The dashed
lines denote the mean values plus/minus their standard errors.

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4. Impact of Increasing CO₂ Concentration on Tidal Dissipation

The previous section demonstrated that tidal amplitudes in the MLT region decrease over time (i.e., with increasing CO_2 concentration), even though they increase below ~82 km. This decrease in tidal amplitudes is likely attributed to enhanced dissipation above ~72 km. This section explores the potential mechanisms responsible for the enhanced dissipation. To the best of our knowledge, tidal dumping in the middle atmosphere is modulated by three primary factors: vertical wavenumber (Forbes and Vincent, 1989; Kogure and Liu, 2021), gravity wave breaking (Forbes and Vincent, 1989; Lu and Fritts, 1993; Meyer, 1999), and meridional shear of zonal wind around 18° in latitude, where tidal zonal wind perturbations peak (Kogure and Liu, 2021; Mayr and Mengel, 2005; McLandress, 2002). This section focuses on changes in vertical wavenumber and gravity wave diffusion, comparing conditions between 2003–2013 and 2050-2061. Discussion of the shear is omitted because the magnitudes of the zonal mean zonal wind shear were found to have decreased in 2050-2061 compared to 2003–2013 (see Figure S1 in supplement), which is favorable for the DW1 tide and reduces a tidal dumping,

4.1. Vertical wavenumbers in 2003–2013 vs. 2050–2061.

Forbes and Vincent (1989) demonstrated that tidal dissipation due to eddy diffusion is approximately proportional to the square of the vertical wavenumber (i.e., inversely proportional to the square of the vertical wavelength). Here, we derived the vertical wavenumbers of the (1,1) mode from the phases shown in Figure S2 (in supplement), using the least-squares method applied over five vertical steps (~9 km) with a step size of ~1.4 km, following the approach of Kogure and Liu (2021). Figure 4a compares the vertical wavenumbers for the periods 2003–2013 (blue) and 2050–2061 (orange). Between ~55 and ~82 km, the vertical wavelengths in 2050–2061 are shorter than those in 2003–2013. Figure 4b shows the difference in vertical wavenumber values between the two periods. A positive difference exceeding one standard error appears throughout most of the ~54-82 km range, with a peak at ~74 km altitude. These results suggest that shorter vertical wavelengths (larger vertical wavenumbers) in 2050–2061 contribute to stronger tidal dissipation and reduced tidal amplitudes, although the change in vertical wavenumbers appears ~15 km below the altitude where the amplitude ratio begins to decline (see Figure 3a).

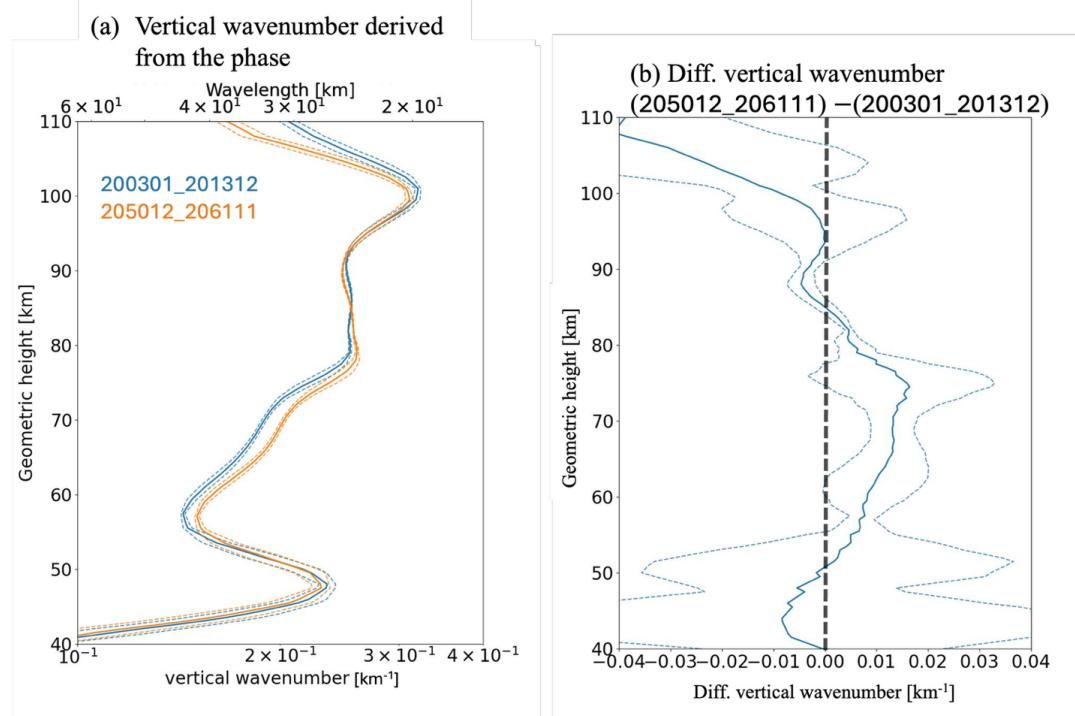



Figure 4. (a) Vertical wavenumbers of DW1 (1,1) mode tides derived from tidal phase values (shown in Figure S2).
The upper and lower x-axes denote vertical wavelength and vertical wavenumber. The orange and blue lines denote
210   vertical wavenumbers averaged over January 2003–December 2013 and December 2050–November 2061,
respectively. (b) Difference in vertical wavenumbers between 2003–2013 and 2050–2060. Dashed lines denote the
standard errors.

Under the WKB approximation and assuming no gravity wave drag, the local tidal vertical wavenumber at a
215   given colatitude, $k_{z(z,\theta)}$, is described as (Forbes and Vincent, 1989; Kogure and Liu, 2021):

$$k_{z(z,\theta)}^2 = \frac{N_{(z,\theta)}^2}{gh'_{(z,\theta)}} - \frac{1}{4H_{(z,\theta)}^2},\tag{5}$$

where $N, g, H,$ and θ represent the buoyancy frequency, gravitational acceleration, scale height, and colatitude,
respectively. h' is the Doppler-shifted equivalent depth, expressed as:

$$h'_{(z,\theta)} = h\left(1 + \frac{u_{(z,\theta)}}{C_0\sin\theta}\right)^4,\tag{6}$$

220   where $h, C_0,$ and u are the equivalent depth (0.69 km for the (1,1) mode) under no background wind, the magnitude
of the migrating diurnal tide phase speed at the equator (~465 ms^{-1}), and the background zonal wind, respectively.
Figure 5a shows the difference in the local vertical wavenumber ($k_{z(z,\theta)2050-2061} - k_{z(z,\theta)2003-2013}$) calculated using
Eq. (2). Below ~92 km, the differences in local wavenumbers at latitudes below ~20°N/S qualitatively match those
derived from tidal phase analysis (Figure 5b), particularly the positive peak within the ~60–80 km range. Since
225   $k_{z(z,\theta)}$ depends on $N, H,$ and h', changes in vertical wavenumber $\delta k_{z(z,\theta)}$ due to variations in these parameters can
be approximated by a Taylor series expansion:

$$\delta k_{z(z,\theta)} \approx \frac{N\delta N}{kgh'} - \frac{N^2\delta h'}{2kgh'^2} + \frac{\delta H}{4kH^3} + O_{(\delta^2)}.\tag{4}$$

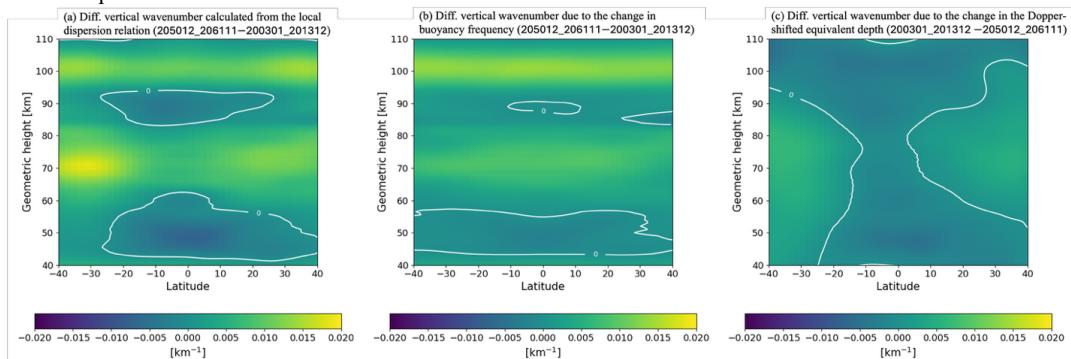
The first and second terms represent the effects of changes in buoyancy frequency and Doppler-shifted equivalent
depth (i.e., zonal mean zonal wind), respectively. The third and fourth terms account for changes in the scale height
230   and high-order terms, which are relatively small and can be neglected (not shown). Figure 5b, c illustrate the effects
of buoyancy frequency and Doppler-shifted equivalent depth, respectively. The buoyancy frequency effect agrees
with the characteristics of $\delta k_{z(z,\theta)}$ below ~92 km, exhibiting a positive peak at ~72 km and two negative peaks near
~48 and ~90 km. Notably, the difference between $\delta k_{z(z,\theta)}$, and the buoyancy frequency effect at ~72 km over the
equatorial region is less than ~5×10^{-4} km^{-1}. Conversely, within the positive $\delta k_{z(z,\theta)}$ region (~60–80 km altitude),
235   the magnitude of the Doppler-shifted effect is approximately three times smaller than that of the buoyancy
frequency effect within 20°N/S, although around the negative $\delta k_{z(z,\theta)}$ layer (~48–90 km), the magnitude of the
Doppler-shifted effect is larger than that of the buoyancy frequency. Therefore, the increase in buoyancy frequency
likely shortens the vertical wavelengths between ~60 and ~80 km altitudes, thereby enhancing tidal dissipation in
the mesosphere.

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Figure 5. (a) Difference in local vertical wavenumber computed from Eq. (2). (b) Effects of buoyancy frequency
changes due to increasing CO$_2$ on the local vertical wavenumber. (c) Same as (b), but showing Doppler-shifted
effects.

245   To further explore the relationship between the increasing CO_2 cooling and the increased buoyancy frequency, we compare temperatures and their vertical gradients between 2003–2013 and 2050–2061. Figure 6 illustrates the differences in zonal mean temperatures (7a) and their vertical gradients (7b), respectively. The negative temperature difference in ~50–70 km is larger (~-9 K at maximum) than that (~-2 K) in ~75–85 km. This strong cooling in ~50–70 km increases atmospheric stability, thereby strengthening buoyancy frequency in ~62–82

250 km.

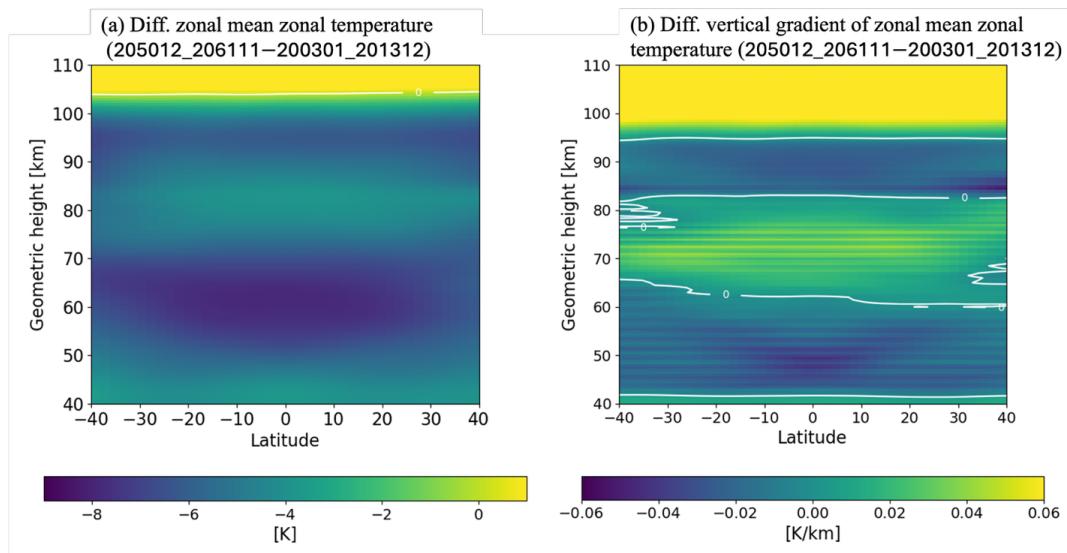



Figure 6. (a) Difference in zonal mean temperature between 2050–2061 and 2003–2013. (b) Same as (a), but showing differences in vertical temperature gradients.

255   Next, we examine the mechanism responsible for the cooler temperatures in the lower to middle mesosphere (~50-70 km). Temperatures in this layer are influenced not only by CO_2 cooling but also by O_3 heating via ultraviolet absorption (Garcia, 2021; Garcia et al., 2019; Jonsson et al., 2004; Lübken et al., 2013). Figure 7 shows the differences in CO_2 and O_3 concentrations, calculated by subtracting the mean values in 2003–2013 from those in 2050–2061. CO_2 concentration increases uniformly by ~200 ppm in ~40–80 km, with the rate of increase

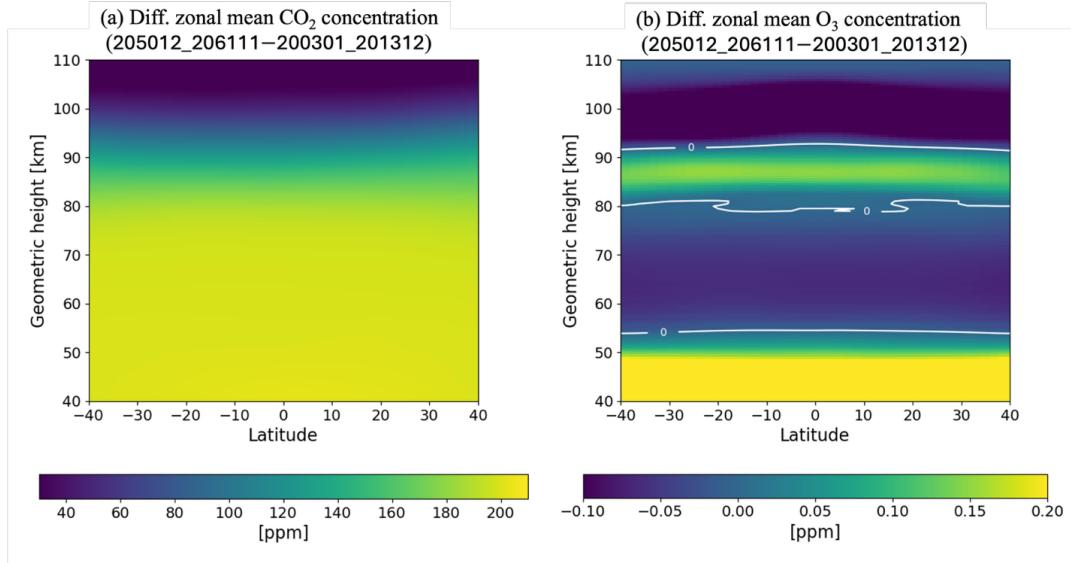
260 sharply declining above ~ 80 km. This indicates that CO_2 is well mixed up to ~80 km, which results in stronger cooling below that altitude. O_3 concentration decreases by up to ~0.05 ppm in ~53-79 km and increases by up to ~0.16 ppm in ~78-92 km during 2050–2061 contributing to cooler temperatures in the lower mesosphere and warmer temperatures in the upper mesosphere.

265   Figure 7. (a) Relative difference in CO_2 concentration. (b) Same as (a), but for the O_3 concentration.

Figure 8 shows vertical profiles of temperature (8a), CO_2(8b), and O_3 concentrations(8c) at 1°N, along with their differences (2050–2061 minus 2003–2013) in panels (8d)–(8f). CO_2 concentrations in Figure 8(b) remain nearly constant up to ~80km and then decrease sharply above that, suggesting that CO_2 concentrations are mixed well up to
270   ~80 km. This vertical feature is seen in its difference (Figure 7a) as aforementioned. These results support the idea that CO_2 cooling is more effective below ~80 km due to the high concentration. Regarding O_3 concentrations in Figure 8(c), the altitudes of the local minimum (~78 km) and maximum (~93 km) shift downward in 2050–2061. This contraction leads to increased O_3 above and decreased O_3 below the ~78 km altitude, resulting in a local maximum (~+0.15 ppm at ~86 km) and minimum (-0.05 ppm at 62 km) in Figure 8f. These levels correspond to a
275   local minimum (~-3 K at ~85 km) and maximum (~-8 K at ~62 km) in temperature decrease between 50 and 90 km in Figure 8d. This correspondence supports that the mesospheric ozone vertical variation contributes to the strengthened stability in the lower and middle mesosphere.

        To summarize this subsection, CO_2 likely induces cooling throughout ~40–80 km than ~80–110 km due to its mixing and transport, while O_3 contributes to cooling in ~53–79 km and warming in ~79–92 km due to the
280   downward shift of the O_3 layer. This downward shift results in a pronounced cold region around ~60 km and intensifies atmospheric stability in ~62–82 km.

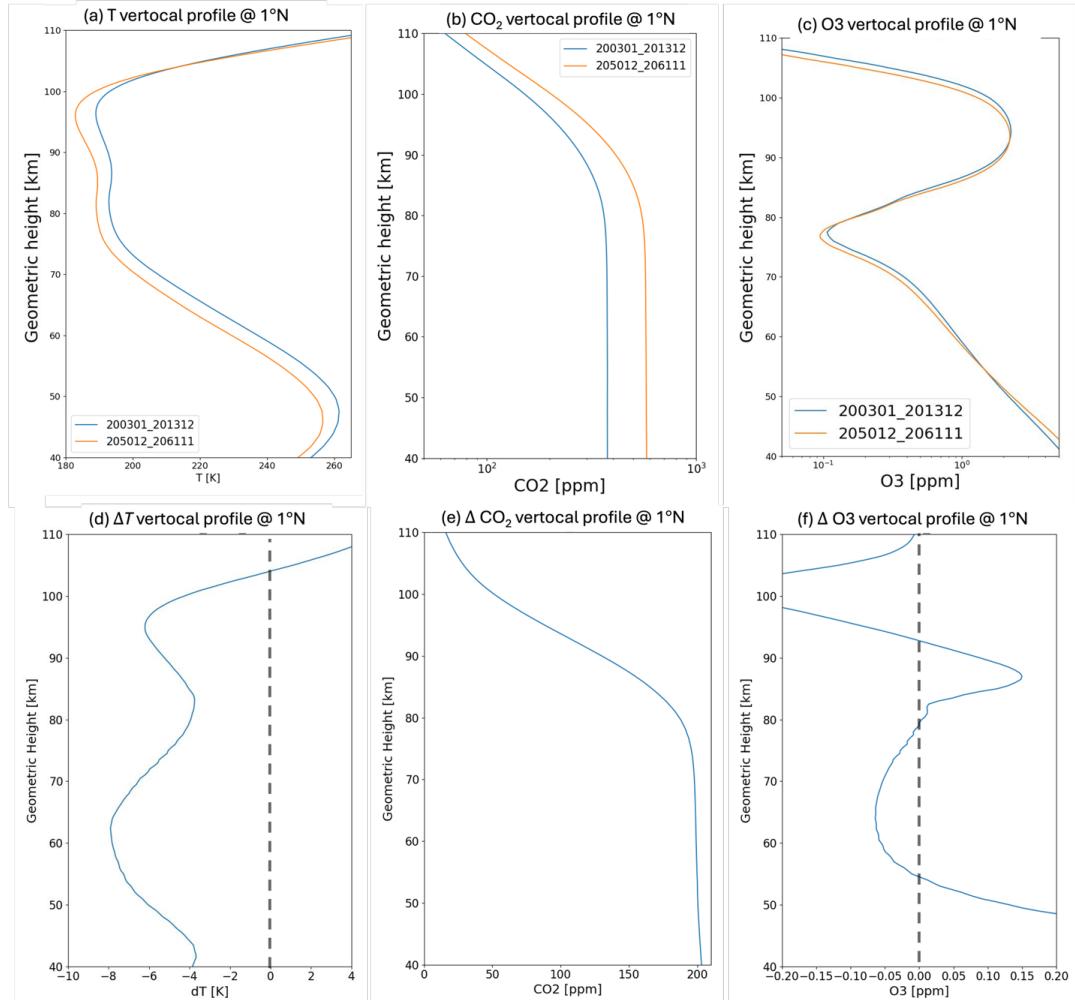



Figure 8. (a) Vertical profile of temperature at 1°N in 200301–201312 (blue) and 205012–20611 (orange). (b) same as (a) but for CO_2 concentration. (c) same as (a) but for O_3 concentration. (d) Difference in temperature at 1°N (205012–20611 minus 200301–201312). (e) same as (d) but for CO_2 concentration. (f) same as (d) but for O_3 concentration.

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4.2. Gravity Wave Drag Parameterization in 2003–2013 vs. 2050‒2061.

According to Meyer (1999), gravity wave breaking interacts with tides through two mechanisms: gravity wave diffusion and the diurnal harmonics of gravity wave drags. Here, we focus on gravity wave diffusion, as the diurnal harmonics of the wave drag were not available due to data storage limitation. It should be noted that the
295 wave drag effect depends on a specific gravity wave scheme used (Mayr et al., 1998; McLandress, 1997; Meyer, 1999; Miyahara and Forbes, 1991), whereas gravity wave diffusion consistently damps tides among the schemes (Meyer, 1999). Saturation and breaking of gravity waves induce diffusion, which dissipates atmospheric waves including tides.

Figure 9 shows the difference in diffusion due to parameterized gravity waves between 2003–2013 and
300 2050–2061. Between 30°N and 30°S, diffusion increases at nearly all altitudes from 40 km to 110 km, with a pronounced peak around 90 km (~1.3 ms^{-2} at maximum), although a localized decrease appears around 10°N. These increases correspond to an ~10% increment from 2003–2013. Since convection is the primary source of gravity waves in equatorial regions, where DW1 tides are concentrated, we examine the difference in the zonal mean precipitation rate. The precipitation rate increases between ~15°S and ~5°N, particularly around ~0°N/S by
305 ~4.5×10^{-9} m $\cdot$ s^{-1}, except for its decrease at ~10°N by ~4.1×10^{-9} m $\cdot$ s^{-1}; the tropical precipitation mostly increases. These changes correspond to the variations in gravity wave diffusion. This increase in tropical precipitation is consistent with findings from previous studies on tropospheric climate change (e.g., Chou and Neelin, 2004; Feng et al., 2019; Lau et al., 2013). In addition, previous numerical modeling studies have reported that increased tropical precipitation intensifies stratospheric GWs and their source activity (Franke et al., 2023;
310 Watanabe et al., 2005). Therefore, increased CO_2 concentrations may strengthen equatorial convection activity, leading to enhanced tropical gravity wave activity. Those enhanced gravity waves are saturated and broken in the MLT layer, which intensifies diffusion and reduce the tidal amplitudes. The GW diffusion in 30-40°N is also enhanced, possibly due to increased frontogenesis, which might also contribute to the tidal dumping.

It should be noted that the strengthened equatorial convection likely leads to the tidal positive trend below
315 ~70 km with the depression in the atmospheric density in the mesosphere, shown in Figures 2 and 3, as well. However, the tidal dissipation associated with the strengthened stability and GW diffusion in the mesosphere could overwhelm the positive trend, resulting in the significant negative trend in the MLT layer.
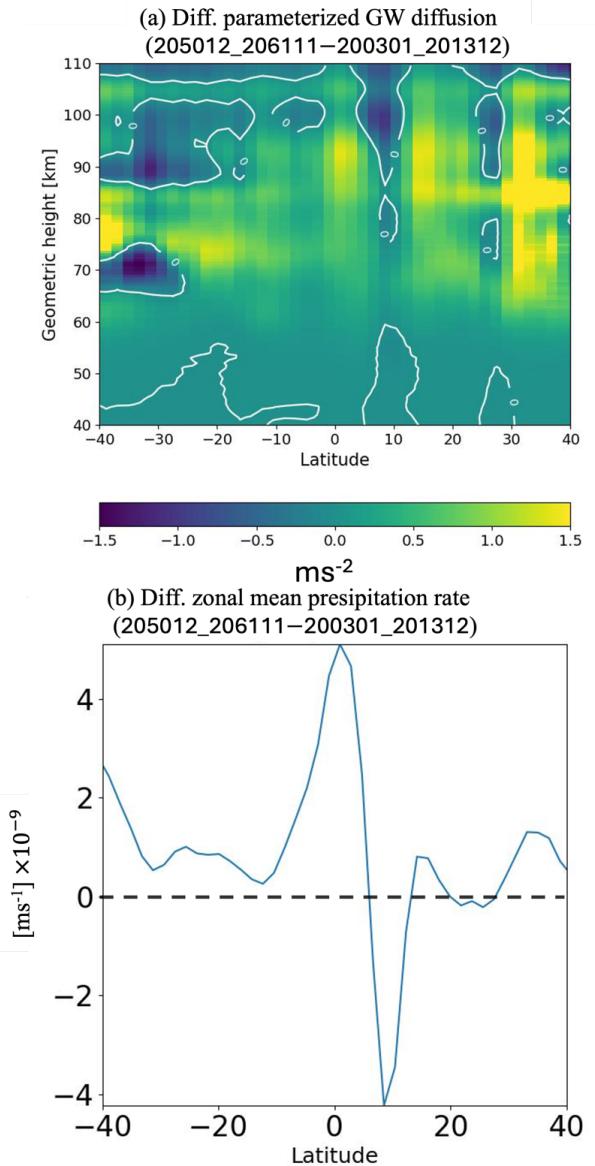
Figure 9. (a) Difference in diffusion due to parameterized gravity waves between 2003–2013 and 2050–2061, shown
as a function of latitude and geometric height. (b) Difference in zonal mean precipitation rate between 2003-2013
and 2050-2061.

320

5. **Conclusion and Discussion.**

325   We examined the response of the DW1 tidal (1,1) mode to increasing CO_2 concentrations using a long-term WACCM-X simulation spanning from January 2000 to July 2069, following the RCP 8.5 scenario. The analysis reveals two significant responses of the DW1 tide to increasing CO_2 levels: (1) Below 70 km altitude, tidal activity increases significantly, likely due to enhanced water vapor and latent heating. Additionally, in 50–70 km, the depression in the atmospheric density likely contributes to the positive trend. The rate of increase is ~10% per 100 years. (2) Above 90 km altitude, tidal activity decreases by ~20% per 100 years, likely due to increased tidal

330   dissipation above ~70 km. It should be noted that the same trends in DW1 tides, background conditions, and gravity wave drag persisted from 2069 August to 2089 December (not shown), even after CO_2 concentrations exceeded the upper limit (720 ppm) of the Fomichev non-LTE CO_2 cooling scheme in the mesosphere. We propose two potential mechanisms contributing to the increased tidal dissipation: a decrease in tidal vertical wavelength and an increase in diffusion due to GW breaking. The shorter vertical wavelength is likely linked to enhanced stability in the

335   mesosphere, as cooling in the lower and middle mesosphere is stronger than in the upper mesosphere. This relatively strong cooling could be attributed to the vertical variations in CO_2 and O_3 concentrations in the mesosphere. CO_2 increases more below ~ 80 km than above, due to its mixing and transport, which induces stronger cooling throughout ~40–80 km than ~80–110 km. Meanwhile, the mesospheric ozone layer shifts downward, leading to decreased concentrations and cooling within ~53–79 km, and increased concentrations and warming within ~79–92

340   km. The combination of CO_2 and O_3 vertical variations intensifies atmospheric stability in ~62–82 km, thereby reducing the tidal vertical wavelengths and amplitudes there. Additionally, the increase in GW diffusion may be attributed to enhanced convective activity, as tropical precipitation is intensified. While this enhanced convective activity likely strengthens tidal activity and contributes to the tidal positive trend below ~70–80 km, the increased tidal dissipation in the mesosphere overwhelms this positive effect, resulting in the significant negative trend in the

345   MLT layer. Taking account into the negative trend clearly increases in the future above ~80 km (see Figure 3a), the GW diffusion might contribute to the tidal damping more than the shorten vertical wavelengths.

        Our findings for the troposphere and stratosphere agree with those of McLandress and Fomichev (2006), whereas our results for the MLT region show an opposite trend. We believe this inconsistency arises from differences in vertical diffusion. In CMAM which simulated the background conditions in McLandress and

350   Fomichev (2006), the lower mesosphere became cooler than the upper mesosphere as CO_2 concentrations increased, consistent with our simulation results (see Figure 10 in Fomichev et al., 2007). However, unlike WACCM-X, which accounts for variations in gravity wave diffusion with increasing CO_2, the tidal linear model used in McLandress and Fomichev (2006) employed a time-independent vertical diffusion coefficient. This likely contributes to the differences in outcomes.

355   Finally, we highlight three major uncertainties in the tidal response to increasing CO_2. The first uncertainty is the increasing CO_2 impact on the stratospheric QBO. Although Wang et al. (2022) suggests that tropospheric global warming increases the frequency of QBO disruption events, the QBO is prescribed with climatology in our future run. Consequently, our simulation does not account for the QBO disruption impacts on the tides, even though such a QBO disruption event has been shown to intensify the (1,1) DW1 tide (Kogure et al, 2021). The second uncertainty

360   is a temperature response in the mesosphere to the CO_2 concentration pathway. Garcia et al. (2019) reported that the cooling rate in the lower and middle mesosphere varies depending on the CO_2 concentration pathway. While the RCP 8.5 scenario leads to strong cooling in the lower and middle mesosphere, consistent with our results, this pronounced cooling was absent under the RCP 6.0 scenario. This suggests that the negative tidal trend in the MLT may vary with the CO_2 concentration pathway. The third uncertainty lies in the effects of GWs on the DW1 tide.

365   The positive trends in tropical GW and DW1 tidal source activity are likely robust, as multiple tropospheric climate studies agree that tropical convective activity will increase with rising CO_2 levels (e.g., Chou and Neelin, 2004; Feng et al., 2019; Lau et al., 2013). Enhanced tropical convection also leads to increased tidal and GW source activity, as reported by McLandress and Fomichev (2006), Franke et al. (2023), and Watanabe et al. (2005). However, the tidal response to parameterized GW momentum deposition depends on the type of GW parameterization scheme used.

370   Parameterizations based on the Lindzen scheme, used in WACCM-X, dissipate tides, while those based on the Hines scheme intensify tides (McLandress 1997; Mayr et al., 1998). Although GW diffusion always damps tides in both schemes, momentum deposition may mitigate the negative tidal trend. Furthermore, the parameterization used in WACCM-X does not account for horizontal GW propagation (e.g., Sato et al., 2009; Kalisch et al., 2014; Kogure et al., 2018; Song and Chun, 2008; Song et al., 2020) or secondary GW generation (e.g., Becker and Vadas, 2018;

375   2020; Vadas and Becker, 2018), both of which can significantly influence momentum deposition and diffusion. H. L. Liu (2021, 2025) pointed out that current gravity wave parameterizations, due to their simplifications,

significantly underestimate diffusion in the MLT layer. Therefore, a more realistic GW parameterization is necessary to accurately assess the impact of increasing CO_2 on the MLT region.

380   Despite these uncertainties, our study reaffirms that increasing CO_2 affects not only the thermal structure, but also the dynamic properties of the MLT region (such as wave activities, diffusion, and circulation diffusion) as previous pointed out (Liu et al., 2020). Our results indicate that the DW1 (1,1) tidal amplitude in the 2050s will be reduced by ~10% in the MLT region (equivalent to a few Kelvin) compared to the 2000s. Current observational techniques are sufficiently accurate to detect temperature variations of this magnitude, suggesting that this negative tidal trend could be confirmed within the next few decades.

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390   **Code and Data availability.**

The long-term future simulation in WACCM-X can be downloaded in https://doi.org/10.5281/zenodo.15189573 (Ma et al., 2025). Our python script for Hough mode decomposition can be obtain in the GitHub repository (Kogure ,2025): https://github.com/masaru-kogure/Hough_Function#.

Competing interests.

395   The authors declare that they have no conflict of interest.

Author contributions.

MK, IS, and HL conceptualized and designed the study. HLL conducted the long-run simulations. MK analyzed the data and wrote the manuscript draft. All the authors reviewed, edited, and contributed to the scientific discussion in the paper.

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