

Thermospheric and Ionospheric Effects by Gravity Waves from the Lower Atmosphere

Han-Li Liu¹, Peter Hjort Lauritzen², Francis Vitt², and Steve Goldhaber³

¹National Center for Atmospheric Research, P. O. Box, 3000, Boulder, CO 80307-3000

²National Center for Atmospheric Research (UCAR)

³NCAR

November 24, 2022

Abstract

A new version of NCAR Whole Atmosphere Community Climate Model with thermosphere/ionosphere extension (WACCM-X) has been developed. The main feature of this version is the species-dependent spectral element (SE) dynamical core, adapted from the standard version for the Community Atmosphere Model (CAM). The SE is on a quasi-uniform cubed sphere grid, eliminating the polar singularity and thus enabling simulations at high-resolutions. Molecular viscosity and diffusion in the horizontal direction are also included. The Conservative Semi-Lagrangian Multi-Tracer Transport Scheme (CSLAM) is employed for the species transport. An efficient regriding scheme based on the Earth System Modeling Framework (ESMF) is used to map fields between the physics mesh and geomagnetic grid. Simulations have been performed at coarse (~ 200 km and 0.25 scale height) and high (~ 25 km and 0.1 scale height) resolutions. The spatial distribution of the resolved gravity waves from the high-resolution simulations compare well with available observations in the middle and upper atmosphere. The forcing by the resolved gravity waves improves the wind climatology in the mesosphere and lower thermosphere in comparison to the coarse resolution simulations with parameterized forcing. It also impacts the thermospheric circulation and compositional structures, as well as thermospheric variability. While larger scale waves are dominant energetically at most latitudes, smaller scale waves contribute significantly to the total momentum flux, especially at mid-high latitudes. The waves in the thermosphere are shown to be strongly modulated by the large-scale wind through Doppler shift and molecular damping, and they cause large neutral atmosphere and plasma perturbations.

Supporting Information for “Thermospheric and Ionospheric Effects by Gravity Waves from the Lower Atmosphere”

H.-L. Liu¹, P. E. Lauritzen², F. Vitt^{1,3} and S. Goldhaber²

¹High Altitude Observatory, National Center for Atmospheric Research, Boulder, Colorado, USA

²Climate and Global Dynamics, National Center for Atmospheric Research, Boulder, Colorado, USA

³Atmospheric Chemistry Observations and Modeling, National Center for Atmospheric Research, Boulder, Colorado, USA

Contents of this file

1. Movie S1

Additional Supporting Information (Files uploaded separately)

1. Captions for Movie S1

Introduction

The animation included is the relative total electron content (TEC) from one day (January 13) of WACCM-X SE NE120/L273 simulation, high pass filtered using the Savitzky-Golay method with period shorter than 2 hours. The file format: QuickTime.

Movie S1. High-pass filtered relative total electron content (TEC) perturbations (unit: percent) with period shorter than 2 hours from WACCM-X SE simulation of January 13.

June 22, 2022, 3:12pm

Abstract

A new version of NCAR Whole Atmosphere Community Climate Model with thermosphere/ionosphere extension (WACCM-X) has been developed. The main feature of this version is the species-dependent spectral element (SE) dynamical core, adapted from the standard version for the Community Atmosphere Model (CAM). The SE is on a quasi-uniform cubed sphere grid, eliminating the polar singularity and thus enabling simulations at high-resolutions. Molecular viscosity and diffusion in the horizontal direction are also included. The Conservative Semi-Lagrangian Multi-Tracer Transport Scheme (CSLAM) is employed for the species transport. An efficient regridding scheme based on the Earth System Modeling Framework (ESMF) is used to map fields between the physics mesh and geomagnetic grid. Simulations have been performed at coarse (~ 200 km and 0.25 scale height) and high (~ 25 km and 0.1 scale height) resolutions. The spatial distribution of the resolved gravity waves from the high-resolution simulations compare well with available observations in the middle and upper atmosphere. The forcing by the resolved gravity waves improves the wind climatology in the mesosphere and lower thermosphere in comparison to the coarse resolution simulations with parameterized forcing. It also impacts the thermospheric circulation and compositional structures, as well as thermospheric variability. While larger scale waves are dominant energetically at most latitudes, smaller scale waves contribute significantly to the total momentum flux, especially at mid-high latitudes. The waves in the thermosphere are shown to be strongly modulated by the large-scale wind through Doppler shift and molecular damping, and they cause large neutral atmosphere and plasma perturbations.

Plain Language Summary

Small scale waves can be excited from daily weather near the Earth surface. These waves, termed gravity waves, can propagate upward and are thought to influence the middle and upper atmospheric regions. Such effects, however, are difficult to directly quantify by observations and numerical modeling due to their small scales and global presence. To address this challenge, we have developed a high-resolution whole atmosphere model (WACCM-X), which extends from the Earth surface to the upper thermosphere, that can partially resolve the small scale waves. The simulated waves are compared with available observations to verify the model results and to examine how these waves are distributed geographically and over altitudes. The forcing of these waves is found to be strong in the thermosphere. It affects the general circulation and the distribution of important atmospheric composition. The simulations also show that wave signatures can be clearly identified in the neutral and the ionized atmosphere, which can have important implications for space weather.

1 Introduction

The thermosphere and ionosphere (TI) system is the nexus of solar radiative and particulate, magnetospheric and lower atmospheric forcing, and affects the bottom-up and top-down coupling processes. As such, the compositional, thermal and dynamical structures of TI and its variation are a focal point of space environment and space weather research. The profound and complex variation of the whole TI system and its interactions with the magnetosphere during solar and geomagnetic storms is of key interests for space weather, and is a central theme of the NASA Diversity, Realize, Integrate, Venture, Educate (DRIVE) Center: the Center for Geospace Storms (CGS) (<https://cgs.jhuapl.edu/>). Meteorological and lower atmosphere forcing have also been shown to affect the TI on weather scales (H.-L. Liu, 2016, and references therein), as well as the TI responses during storm time (Pedatella & Liu, 2018). The lower atmosphere can also affect the TI system from subseasonal-to-seasonal scales (Gasperini et al., 2017, 2020; Richter et al., 2022) to variations associated with anthropogenic climate change (Roble

63 & Dickinson, 1989; Emmert et al., 2004, 2008; Solomon et al., 2018). The variability in
 64 the coupled TI system strongly influences the operation of Low-Earth Orbiting (LEO)
 65 satellites and space station, GPS navigation and radio communication. The dynamics
 66 of the middle atmosphere and TI determines the upward transport and escape of atomic
 67 hydrogen (Catling & Kasting, 2017; Jones Jr. et al., 2020) and downward transport of
 68 NO_x (Randall et al., 2006, 2009), which are species important for the long-term evolu-
 69 tion of the Earth atmosphere and the climate variability of the middle and lower atmo-
 70 sphere, respectively.

71 Mesoscale processes, in particular atmospheric gravity waves (GWs), are thought
 72 to play an important role in coupling of the lower atmosphere with the space environ-
 73 ment. While GWs are increasingly damped by the strong molecular viscosity in the ther-
 74 mosphere, waves with large vertical propagating speed can still penetrate to high alti-
 75 tudes (Pitteway & Hines, 1963). This is evidenced in observational studies relating TEC
 76 perturbations to strong tropospheric deep convective storms (e.g. Nishioka et al., 2013;
 77 Azeem et al., 2015, 2018). Limited satellite observations have shed light on the distri-
 78 bution and seasonal variation of GWs around 250 km, though it is not clear how they
 79 are related to the sources in the lower atmosphere (Park et al., 2014; Forbes et al., 2016;
 80 H. Liu et al., 2017). For the ionosphere, GWs can induce traveling ionospheric distur-
 81 bances (TIDs), and there have been decades of study on their role in seeding ionospheric
 82 irregularities (e.g. Huang & Kelley, 1996a, 1996b; Hysell et al., 1990; Krall et al., 2013;
 83 Kelley et al., 1981; McClure et al., 1998; Retterer & Roddy, 2014). In recent years, the-
 84oretical and high-resolution numerical simulations have advanced our understanding of
 85 the GWs in the thermosphere. Secondary GWs from the dissipation of primary GWs
 86 can have large spatial scales and fast propagating speeds, and can thus play an impor-
 87 tant role in the upper atmosphere dynamics (Vadas & Fritts, 2001; Vadas, 2007; Vadas
 88 & Liu, 2013; Vadas & Becker, 2019; Becker & Vadas, 2020). It is also demonstrated that
 89 GWs can indeed seed equatorial plasma bubbles (EPBs) (Huba & Liu, 2020).

90 The need to understand the TI system in the context of its interactions with the
 91 lower atmosphere has motivated the development of numerical models spanning the whole
 92 atmosphere region, from Earth surface to the upper thermosphere, that can simulate the
 93 radiative, chemical, dynamical, and electrodynamical processes. A review of whole at-
 94 mosphere modeling was given by Akmaev (2011), and model capabilities have advanced
 95 further since then, including enhanced thermospheric and ionospheric physics (Jin et al.,
 96 2011; Verronen et al., 2016; H.-L. Liu et al., 2018; Borchert et al., 2019; Wu et al., 2021)
 97 and whole atmosphere data assimilation (Wang et al., 2011; Pedatella, Raeder, et al.,
 98 2014; Pedatella et al., 2020). It has been recognized, however, that one of the major sources
 99 of bias in whole atmosphere models is the uncertainties associated with the parameter-
 100 ization of GW effects on forcing and transport (Pedatella, Fuller-Rowell, et al., 2014),
 101 which are necessitated by the limited model resolution. The poor model resolution also
 102 limit the whole atmosphere capabilities in resolving other TI effects by GWs (e.g. TIDs
 103 and ionospheric dynamo), and mesoscale processes in general. This need, along with the
 104 improved numerical algorithms and increasing computational power, has motivated and
 105 enabled the development of the high-resolution capability of whole atmosphere models
 106 (Watanabe & Miyahara, 2009; H.-L. Liu et al., 2014; Miyoshi et al., 2018; Becker & Vadas,
 107 2018; Becker et al., 2022; Okui et al., 2022).

108 In the current study, we have developed the high resolution capability for the NCAR
 109 Whole Atmosphere Community Climate Model with thermosphere/ionosphere extension
 110 (WACCM-X), which makes it possible to resolve the meso- α range (200–2000km). With
 111 the model capability in solving interactive chemistry, transport, and ionospheric electro-
 112 dynamics, we will examine the GW distribution and variation from the lower atmosphere
 113 to the upper thermosphere, the wave effects on the thermospheric circulation and com-
 114 position, the scale dependence of wave contribution to the energy and momentum bud-
 115 get, and TI perturbations caused by the waves.

2 Model Description

WACCM-X is one of the atmosphere components in the NCAR Community Earth System Model (CESM). The thermospheric and ionospheric physics of WACCM-X is described in H.-L. Liu et al. (2018). The lower and middle atmosphere physics packages used in WACCM-X have recently been updated to WACCM version 6 (WACCM6) (Gettelman et al., 2019). Earlier versions of WACCM-X have been using finite volume dynamical core on a traditional latitude and longitude grid. One disadvantage of the latitude-longitude grid is the polar singularity: the size of the grid decreases toward 0 when approaching the pole. Polar filtering or smoothing has to be applied to maintain numerical stability, which affects the fidelity of the physics in the polar region and degrades the model performance at higher spatial resolutions. A new dynamical core option in CESM atmosphere components is the spectral element (SE) core, which employs a quasi-uniform cubed sphere grid (Lauritzen et al., 2018). This addresses the polar singularity issue, and enables simulations at much higher horizontal resolutions. The high-resolution capability of the SE dynamical core and its ability to resolve gravity waves was demonstrated in an experimental version of WACCM with specified chemistry (up to ~ 140 km) (H.-L. Liu et al., 2014).

The SE dynamical core has recently been adapted for WACCM-X, by taking into consideration the species-dependence of mean molecular weight, dry air gas constant, and specific heats, which are necessary to properly resolve the thermosphere dynamics. Formulations of these quantities used are described in H.-L. Liu et al. (2018), and similar formulations are used here. It is noted that in the SE dynamical core, the thermodynamics quantity used is temperature, not potential temperature. There is thus no need to correct for variable κ (the ratio between gas constant R and specific heat at constant pressure c_p). The implementation of species dependent thermodynamics uses the generalized thermodynamic infrastructure for moist air containing any number of forms of water (Lauritzen et al., 2018). This infrastructure has been extended to species dependent dry air so that the user specifies the major species via name list and the model automatically adapts. The details are given in Appendix A.

In addition to the modifications of the dynamical core thermodynamics, molecular viscosity and thermal conductivity (which previously only has been represented in the physics package in the vertical) is now also represented in the horizontal in the SE dynamical core. The formal equations are given in Appendix B. Again we highlight that similarly to the thermodynamic infrastructure, the molecular viscosity and thermal conductivity coefficients are computed by a common module shared between physics and dynamics to ensure consistency and flexibility. In addition to the physical damping, artificial viscosity needs to be applied to maintain numerical stability, as noted in the Appendix C. This artificially damps gravity waves, and the simulated waves discussed in the paper are therefore likely to be weaker than reality.

For uniform resolution applications the NCAR SE dynamical core has the option to use an accelerated transport scheme called CSLAM (Lauritzen et al., 2017, Conservative Semi-Lagrangian Multi-tracer scheme) and coupling to physics using a finite-volume physics grid (Herrington, Lauritzen, Taylor, et al., 2019; Herrington, Lauritzen, Reed, et al., 2019). This option provides more accurate and efficient (if enough tracers as is the case for WACCM-X) tracer transport and the use of a finite-volume physics grid alleviates spurious noise near element corners/edges (Herrington, Lauritzen, Taylor, et al., 2019).

In WACCM-X the ionospheric electric dynamo is computed in a modified magnetic apex coordinate system (H.-L. Liu et al., 2018; Richmond, 1995). The F-region O^+ transport equation is solved on its own latitude and longitude grid. It is thus necessary to perform efficient run-time communications for state variables among different grids. This is achieved via Earth System Modeling Framework (ESMF, Theurich et al., 2016, <https://>

168 earthsystemmodeling.org/) regriding operations. The WACCM-X data originate in
 169 the atmospheric physics layer of CESM and an unstructured ESMF mesh is created from
 170 the normally column-based physics by reading in a pre-computed ESMF mesh file. This
 171 allows all physics fields to be treated as gridded data. The dynamo and ion transport
 172 grids are rectangular (geomagnetic and geographic latitude/longitude, respectively) and
 173 ESMF objects for those are created dynamically at model initialization time. The dy-
 174 namo grid is recreated yearly to account for changes in the geomagnetic main field. ESMF
 175 regriding operations are then used to transmit data between these grids.

176 Two model configurations are used for this study. For the high-resolution simula-
 177 tions, the horizontal configuration of the cubed sphere is NE120, corresponding to a quasi-
 178 uniform resolution of ~ 25 km. The vertical resolution is 0.1 scale height in most of the
 179 middle and upper atmosphere. In the top 3 scale heights, the vertical resolution transi-
 180 tions to 0.25 scale height, because most of the waves with short vertical wavelengths
 181 are gone due to strong molecular damping. There are a total of 273 levels (L273). As
 182 a comparison, WACCM-X simulations have also been performed with NE16 horizontal
 183 resolution (~ 200 km) and 0.25 scale height vertical resolution for the middle and upper
 184 atmosphere (130 levels, L130). For both configurations, the model top is at 4×10^{-10} hPa
 185 (~ 600 km). The solar radio flux at 10.7 cm, a proxy used to parameterize solar extreme
 186 ultraviolet (EUV) irradiance, is set to 120 solar flux unit (SFU). The geomagnetic in-
 187 dex Kp, used to drive Heelis empirical model for high latitude electric potential spec-
 188 ification, is set to 0.33. Gravity wave parameterization scheme (Gettelman et al., 2019,
 189 and references therein) is used in the WACCM-X NE16/L130 simulations, but is turned
 190 off in WACCM-X NE120/L273.

191 3 Results

192 3.1 Distribution of Resolved Gravity Waves

193 Gravity wave activity level in the high resolution simulations is quantified as fol-
 194 lows: neutral and ionospheric quantities, such as winds and temperature, electron den-
 195 sity and total electron content (TEC), are first high-pass filtered in the zonal direction,
 196 retaining perturbations with zonal scales less than 2000 km. The standard deviation of
 197 these quantities are then computed in $2.5^\circ \times 2.5^\circ$ latitude/longitude bins for each out-
 198 put time step. The standard deviation is used in this study to characterize the longitude,
 199 latitude, altitude and time dependence of the gravity wave activity.

200 Figure 1 shows the zonally averaged standard deviation of zonal, meridional and
 201 vertical winds and temperature for January. It is first noted that in the stratosphere, meso-
 202 sphere and lower thermosphere the latitude/height distribution show the same morphol-
 203 ogy as previous observational studies (e.g., Ern et al., 2011; John & Kumar, 2012; Ern
 204 et al., 2018; Geller et al., 2013): (1) the winter hemisphere maximum is located at higher
 205 latitudes ($50\text{-}60^\circ\text{N}$); (2) the summer hemisphere maximum is located at $\sim 20^\circ\text{S}$ in the
 206 stratosphere and lower mesosphere, and shifts to higher latitudes above, coinciding with
 207 the large eastward wind shear. These features are also well captured in high-resolution
 208 WACCM simulations (H.-L. Liu et al., 2014; H.-L. Liu, 2016).

209 The standard deviations of zonal and meridional winds have similar latitude/height
 210 structure and similar magnitude. In the winter hemisphere (NH), their maximum val-
 211 ues ($20\text{-}30\text{ms}^{-1}$) extend from the middle to high latitudes and 100–200 km, while in the
 212 summer hemisphere (SH) their maximum values are at mid-latitudes and located in the
 213 mesosphere and lower thermosphere (MLT) region. This difference is caused by the hemi-
 214 spheric/seasonal dependence of the molecular damping, which depends sensitively on the
 215 thermosphere temperature and is thus stronger in the summer hemisphere.

216 The maximum values of the vertical wind standard deviation, $\sim 10\text{ms}^{-1}$, are also
 217 found at middle and high latitudes, but at higher altitudes than the horizontal winds

218 (around 300km). This is likely due to the dominance of higher frequency gravity waves
 219 at higher altitudes, because only waves with the largest vertical propagating speeds and
 220 longest vertical wavelengths can survive the molecular damping. Furthermore, the buoy-
 221 ancy frequency increases with altitude. Since the ratio of the vertical and horizontal wind
 222 perturbations is approximately proportional to the ratio between wave frequency and
 223 buoyancy frequency according to the polarization relation, vertical wind perturbations
 224 at upper thermosphere can grow further with altitude even when the horizontal wind
 225 perturbations start to decrease.

226 The temperature perturbations have the maximum standard deviation values (up
 227 to 30 K) spanning 100 and 200 km in NH, and at around ~ 150 km in SH. These large
 228 values are again located at middle to high latitudes.

229 The longitude/latitude dependence of temperature perturbations, averaged over
 230 4 universal times (0, 6, 12 and 18 hours), are shown in Figure 2 for 4 levels (78, 1, 10^{-4} ,
 231 and 10^{-7} hPa, corresponding to approximately the tropopause, stratosphere/stratopause,
 232 lower thermosphere, and upper thermosphere). At the tropopause height, strong activ-
 233 ities are seen over major mountain ranges and plateaus, and over the Western Pacific
 234 warm pool. The distribution changes with altitude. For example, the strong orographic
 235 waves in both hemispheres disappear or are much weakened. In the summer/southern
 236 hemisphere, the zonal wind reversal at the tropopause and lower stratosphere form crit-
 237 ical layers for the orographic gravity waves. Although the zonal mean zonal wind in the
 238 winter/northern stratosphere and mesosphere is eastward, mid-latitude large-scale zonal
 239 wind at a specific longitude can be westward or 0 associated with the quasi-stationary
 240 planetary wave. For example, at $\sim 35^\circ\text{N}$ the westward tilting easterly (westward wind)
 241 phase in the stratosphere and mesosphere extends across almost the entire latitude cir-
 242 cle except between $60\text{--}95^\circ\text{W}$. As such, orographic gravity waves over Tibetan plateau and
 243 the Rocky Mountains are removed by critical layer filtering.

244 By comparing the standard deviation values at 1 and 78 hPa, it is seen that the
 245 wave activities over the Western Pacific is still prominent (over 3 K), but the peak has
 246 shifted southward by about 10° latitude. There is actually a band of activities centered
 247 around 20°S , with large standard deviation values (2-3 K) over the Central Pacific, West-
 248 ern South Atlantic/East Coast of Brazil, and Western Indian Ocean/Madagascar. These
 249 are likely convectively generated gravity waves that propagate southward. In the win-
 250 ter/north hemisphere, large activities are found poleward of 45°N , where the large-scale
 251 zonal wind are predominantly eastward at all longitudes. Peak standard deviation val-
 252 ues extend from the eastern part of North America, across the north Atlantic, to west-
 253 ern Europe, with magnitude of over 3K. Activities are also strong (>2 K) over the rest
 254 of Eurasia poleward of 45°N . By comparing to the stream function of the flow at that
 255 level, it is seen that these large activity regions correspond to the strongest eastward strato-
 256 spheric jet as well as its exit region. The longitude and latitude distribution of wave ac-
 257 tivities in both hemispheres is in good agreement with the stratospheric gravity waves
 258 during January obtained from SABER measurements (Ern et al., 2011).

259 In the lower thermosphere, the large activities in the winter/northern hemisphere
 260 are still located at high latitudes, and the band of activities in the summer/southern hemi-
 261 sphere shifts further southward (as also seen in Figure 1), peaking at around 45°S . The
 262 time averaged longitude dependence of the wave activity in the NH is similar to that at
 263 the stratosphere, with peak values (over 10 K) extending from eastern North America
 264 to central Eurasia. The spatial distribution appears to be more uniform in comparison
 265 with the stratosphere. This results from both the horizontal propagation of the waves,
 266 and stronger time dependence due to tidal waves (and the time averaging) as will be dis-
 267 cussed later. An equatorial band is also seen at the lower thermosphere, with magnitude
 268 up to ~ 6 K. These waves may originate from relatively weak convective activities in the
 269 troposphere. They become more prominent and distinguished at the lower thermosphere

270 resulting from amplitude growth and dispersion of the waves propagating in the merid-
 271 ional direction.

272 In the upper thermosphere, the longitude dependence of the time averaged wave
 273 activity in the NH display a pattern similar to that in the lower thermosphere, with large
 274 values (up to 12 K) from eastern North America to central Eurasia between 45° – 60° N.
 275 The SH activity band shifts further southward, centered around 55° S, with magnitude
 276 (5 K) smaller than the NH due to stronger molecular damping. However, the maximum
 277 standard deviation values in both hemisphere are found near the geomagnetic poles, with
 278 values up to 17 K in the NH and 13 K in the SH. This is consistent with the thermo-
 279 spheric wave distribution obtained from CHAMP and GOCE measurements (Park et al.,
 280 2014; H. Liu et al., 2017). The corresponding maximum standard deviation of zonal wind
 281 perturbations at these locations have similar values to that of the temperature pertur-
 282 bations (17/13 K in NH and SH), which are comparable to the values from GOCE (H. Liu
 283 et al., 2017) (square amplitude of $100 \text{ m}^2\text{s}^{-2}$). The two peaks are likely caused by the
 284 reduced ion drag, when the gravity wave phase lines become more aligned with the field
 285 lines.

286 Gravity waves at higher altitudes display strong local time dependence, in contrast
 287 to the stratosphere (Figure 3). This is associated with the increasingly strong tidal mo-
 288 tion and day-night difference at higher altitudes. At 1.1×10^{-4} hPa, this is most clear
 289 in the NH at mid to high latitudes, with stronger activities around LT 18 hour and 6 hour,
 290 when the large-scale winds are eastward. Diurnal variation is more prominent in the up-
 291 per thermosphere: strong activities are found around LT 9 hour in the NH and LT 6 hour
 292 in the SH. The large-scale winds at these times are westward and poleward, and the grav-
 293 ity waves propagation is opposite to the winds. This sensitive dependence on the wind
 294 direction is because of the large molecular damping. The molecular damping is inversely
 295 proportional to the square of the vertical wavelength. Because the vertical wavelength
 296 decreases when propagating in the same direction of the wind due to doppler shift, these
 297 waves are more severely damped than the ones that propagate against the wind. It is
 298 worth noting that the peak wave activities at different altitudes are not always co-located.
 299 This implies that the waves peaking at different altitudes may originate from different
 300 sources, different spectral portion of the wave sources, and/or secondary generation of
 301 gravity waves. The local time dependence also underscores the significance of large-scale
 302 wind modulation.

303 **3.2 Gravity wave forcing and impact on thermospheric circulation, com-** 304 **position and variability**

305 The zonal mean zonal and meridional forcings by gravity waves with zonal scales
 306 less than 2000 km are calculated from the vertical divergence of the vertical fluxes of zonal
 307 and meridional momentum, respectively (Figure 4). By comparing the zonal forcing with
 308 the zonal mean zonal wind, it is seen that the former is responsible for the zonal wind
 309 reversal in the MLT. Further, there is a clear hemispheric/seasonal asymmetry of the re-
 310 versal, with the eastward summer reversal stronger and located at lower altitudes (~ 90
 311 km) and the westward winter reversal weaker at higher altitudes (above 100 km). This
 312 differs from WACCM-X simulation results obtained using parameterized gravity wave
 313 forcing and is in better agreement with observations (Swinbank & Ortland, 2003; Sto-
 314 ber et al., 2021). Above the primary zonal wind reversal, the zonal gravity wave forc-
 315 ing changes direction again (to westward/eastward in the summer/winter hemisphere)
 316 within a rather shallow region (between about 100–120 km). This results from the dis-
 317 sipation of gravity waves that filter through the primary wind reversal in the MLT. This
 318 feature was qualitatively captured by parameterized gravity wave forcing (H.-L. Liu &
 319 Roble, 2002; H.-L. Liu, 2007), and it is responsible for driving a return circulation (winter-
 320 to-summer) between the summer-to-winter circulations in the MLT and the upper ther-
 321 mosphere.

322 In the upper thermosphere, the predominant wind is westward/eastward in the sum-
 323 mer/winter hemisphere and summer-to-winter in the meridional direction. This is driven
 324 primarily by the differential heating. The zonal mean zonal forcing by gravity waves is
 325 eastward in both hemisphere, and the maximum value in the winter hemisphere (over
 326 $100 \text{ ms}^{-1}\text{d}^{-1}$) is larger than that in the summer hemisphere ($\sim 50 \text{ ms}^{-1}\text{d}^{-1}$). They are
 327 located at mid to high latitudes and 200–300 km altitudes in both hemispheres. While
 328 the eastward forcing is opposite to the mean zonal wind in the summer thermosphere,
 329 it is in the same direction as that in the winter thermosphere. Recall that the gravity
 330 wave activities are the strongest in local morning (Figure 3, LT 9 hour in the NH and
 331 6 hour in the SH), and propagate eastward relative to the large-scale wind. The dissi-
 332 pation of these waves produces a net eastward forcing. The dominance of the eastward
 333 waves in the upper atmosphere may arise from the anisotropy of the source spectrum
 334 and/or filtering by the wind system. In earlier parameterization studies, similar anisotropy
 335 (with stronger eastward waves) was found necessary to reproduce the hemispheric/seasonal
 336 asymmetry of the wind reversal level in the MLT (H.-L. Liu & Roble, 2002). The zonal
 337 mean meridional forcing becomes large in the thermosphere, with magnitudes and hemi-
 338 sphere and altitude distribution similar to the zonal forcing. The direction of the forc-
 339 ing is from winter to summer, against the summer-to-winter meridional circulation.

340 The zonal and meridional forcing by gravity waves have important implications for
 341 the thermospheric circulation. In the NH, the Coriolis force associated with the eastward
 342 forcing is equatorward, thus it partially offsets the westward forcing by large-scale waves
 343 (Figure 4, mainly from migrating tides) and tends to weaken the summer-to-winter cir-
 344 culation. This reinforces the effect by the meridional forcing. On the other hand, the equa-
 345 torward Coriolis force associated with the eastward forcing in the SH tends to strengthen
 346 the summer-to-winter circulation, thus offsetting the effect by the meridional forcing. By
 347 using NE16 as a reference, the zonal mean zonal wind at 200 km and $50\text{--}60^\circ\text{N}$ from NE120
 348 is eastward and faster by $\sim 15 \text{ ms}^{-1}$, which results in a southward Coriolis forcing of ~ 150
 349 $\text{ms}^{-1}\text{d}^{-1}$. This is comparable to the southward gravity wave forcing of $100 \text{ ms}^{-1}\text{d}^{-1}$.
 350 On the other hand, the zonal mean zonal wind at 200 km and $50\text{--}60^\circ\text{S}$ from NE120 is
 351 westward and over 7 ms^{-1} slower, leading to a northward Coriolis forcing of $\sim 70 \text{ ms}^{-1}\text{d}^{-1}$.
 352 This is stronger than the southward gravity wave forcing ($30\text{--}40 \text{ ms}^{-1}\text{d}^{-1}$) at that lati-
 353 tude range. This hemispheric difference in forcing is reflected in the hemispheric differ-
 354 ences zonal mean meridional and vertical winds (Figure 5). The figure shows the aver-
 355 ages around 60° latitudes in both hemispheres, where the gravity wave forcing in NE120
 356 is strong. Both meridional and vertical winds are weaker in the NH and stronger in the
 357 SH above ~ 200 km. It is noted that the gravity wave forcing in the MLT tends to en-
 358 hance the summer to winter circulation. The thermospheric gravity wave forcing in the
 359 NH (winter) hemisphere therefore offsets the effect by the MLT waves.

360 The change of the meridional/vertical circulation affects the thermospheric com-
 361 positional structures. In the winter thermosphere, the downward circulation tends to bring
 362 down the atomic oxygen (O) rich atmosphere and increase its mixing ratio. The slower
 363 circulation, due to the compounded effect of Coriolis force associated with eastward forc-
 364 ing and the opposing meridional forcing, therefore leads to the decrease of the O mix-
 365 ing ratio. On the other hand, the upward circulation in the summer thermosphere has
 366 the opposite effect (reduce O mixing ratio). The Coriolis force therefore tends to reduce
 367 the O mixing ratio, while the opposing meridional forcing tends to increase the O mix-
 368 ing ratio. Figure 6 shows the zonal mean, column integrated O/N_2 (simply referred to
 369 as O/N_2 thereafter), a proxy often used for thermospheric composition in measurements
 370 (Strickland et al., 1995; Meier, 2021), from simulations with high and coarse resolutions.
 371 The O/N_2 from the coarse resolution simulation is generally larger than observations,
 372 especially at higher latitudes in the winter hemisphere. This ratio in the high resolution
 373 simulation shows an overall decrease, in better agreement with observations. The largest
 374 decrease in the winter hemisphere (from 2.5 to 1.6 in the polar region) is consistent with
 375 the compounded effect from both zonal and meridional gravity wave forcing. Despite the

376 enhanced upward circulation in the SH, O/N_2 still decreases, albeit with a smaller mag-
 377 nitude than in the NH (by ~ 0.1). This is likely due to the enhanced effective diffusion
 378 by gravity waves (H.-L. Liu, 2021), which also tends to decrease O/N_2 . As shown by Qian
 379 et al. (2009), adding eddy diffusion at the lower boundary of TIE-GCM could correct
 380 the overestimation O/N_2 in the model and achieve better agreement with TIMED/GUVI
 381 measurements. The high-resolution simulations therefore elucidate the processes by which
 382 gravity waves can affect the O/N_2 , namely both circulation and diffusion. The improved
 383 agreement of O/N_2 also serves a validation of the resolved gravity waves.

384 The changes in the major species and the transport cause changes in minor species.
 385 For example, the global mean NO is larger in the high resolution simulation than that
 386 from the coarse resolution simulation above ~ 70 km. The increase in the thermosphere
 387 is likely due to the larger abundance of N_2 , and the increase in the mesosphere proba-
 388 bly results from enhanced transport by the resolved waves (H.-L. Liu, 2021). Because
 389 of this NO increase, NO cooling rate also increases. For example, the peak NO cooling
 390 rate at $60^\circ S$ and ~ 130 km from the high resolution simulation is $\sim 35\%$ larger than the
 391 coarse resolution simulation (Figure 7). Another important energetics quantity for the
 392 TI system is the Joule heating. The vertical profiles of Joule heating at the same loca-
 393 tion from the high and coarse resolution simulations are shown in the same figure. The
 394 peak value at ~ 120 km from the high resolution simulation is $\sim 33\%$ larger than the lower
 395 resolution results. The increase of Joule heating is likely due to the smaller scale vari-
 396 ations in ion drifts and winds, which are important for producing Joule heating (Codrescu
 397 et al., 1995). As will be shown in Section 3.4, smaller scale perturbations of ion drifts
 398 are excited by gravity waves. The increases of the NO cooling and Joule heating in high
 399 resolution simulations are seen at all latitudes (Figure 7).

400 The mean circulation varies strongly from day to day. Figure 8 shows the zonal mean
 401 and daily averaged meridional and vertical winds at $60.5^\circ N$ near 250 km from 13 to 31
 402 January from both the high-resolution (solid) and coarse-resolution (dotted) simulations.
 403 The ranges of the meridional wind variation during this time period are similar between
 404 the two simulations ($10\text{--}18\text{ ms}^{-1}$), but the day-to-day variation is stronger in the high-
 405 resolution simulation. The difference in day-to-day variation of the mean vertical wind
 406 is more significant: between 0 and -0.8 ms^{-1} in the high-resolution simulation and -0.35
 407 and -0.55 ms^{-1} in the coarse-resolution simulation. The large day-to-day variability of
 408 the circulation results from the variability of the driving force by waves. This is demon-
 409 strated in Figure 9 by the mean zonal and meridional forcing by gravity waves (with zonal
 410 scales less than 2000 km). Both vary between 0 and $200\text{ ms}^{-1}\text{d}^{-1}$.

411 Tides and tidal variability are affected by gravity waves. Figure 10 compares the
 412 mean and standard deviation of the amplitudes of three major tides (migrating diurnal
 413 (DW1) and semi-diurnal (SW2), and non-migrating diurnal, eastward propagating wavenum-
 414 ber 3 (DE3)) from the high-resolution and coarse resolution simulations. The latitude/height
 415 structures of the tides are similar from the simulations, and are consistent with their cli-
 416 matology. The amplitudes of DW1 and DW3 from the simulations are also comparable,
 417 while SW2 in the high-resolution simulation has larger amplitude, especially its peak value
 418 in the winter lower thermosphere. The standard deviation values of all three components,
 419 on the other hand, are larger throughout the middle and upper atmosphere in the high-
 420 resolution simulations.

421 3.3 Scale dependence of wave power and momentum flux

422 With the horizontal resolution of ~ 25 km, the model can effectively resolve grav-
 423 ity waves with horizontal wavelengths longer than ~ 200 km. The meso- β scales are poorly
 424 resolved or unresolved, but they can still have important contribution to the momentum
 425 budget up to the lower thermosphere (H.-L. Liu, 2019). Here we examine the scale de-
 426 pendence of the kinetic energy and momentum flux in the thermosphere and the con-

427 tributions from waves of different scales using the same method proposed by H.-L. Liu
428 (2019).

429 As shown in Figure 11, the zonal wavenumber power spectral density (PSD) of the
430 zonal wind and spectra of vertical flux of zonal momentum still follow power-law in the
431 thermosphere. It is also seen the transition from steeper slopes at large scales to shall-
432 low slopes at smaller scales occurs at larger wavenumbers (between 10 and 20) in the ther-
433 mosphere. This indicates the relatively lower level of gravity wave activity, probably due
434 to increasing molecular damping.

435 Figure 12 shows the contributions to the PSD of the zonal wind and the momen-
436 tum flux spectrum within the zonal scale of 30–2000 km, by waves with zonal scales be-
437 tween 300–2000 km and 30–300 km. The waves within the scale ranges between 300 and
438 2000 km are fully resolved, while waves with scales less than 200 km are under- or un-
439 resolved. One of the considerations for looking at 30–300 km is that it is the measure-
440 ment range of the upcoming NASA Atmospheric Waves Experiment (AWE) mission (Taylor
441 et al., 2017). There are clear differences between the spectral compositions in the ther-
442 mosphere (> 100 km) and below, and between PSDs and momentum flux spectra. The
443 larger scale waves are more prominent in the thermosphere, and they contribute more
444 to PSDs due to their steeper spectral slopes. Between 100–300 km, waves with zonal scales
445 between 300 and 2000 km contribute to over 75% of the total zonal kinetic energy (be-
446 tween 30 and 2000 km) at low and mid-latitudes, and down to 40% (NH) and 55% (SH)
447 at high latitudes. Below 100 km the contribution can drop to 40% around 20°S and 20%
448 at $40\text{--}50^{\circ}\text{N}$. The percentage contribution of this scale range to the momentum flux shows
449 a similar spatial variation, but with lower values: 45–60% between 100–300 km and 40°S –
450 40°N , and 15–50% at higher latitudes; below 100 km less than 10% at mid to high lat-
451 itudes in both hemispheres and up to 60% over the equator. The percentage contribu-
452 tions by the 30–300 km are the residual of those by 300–2000 km, but they are shown in
453 Figure 12 for clarity. In the middle atmosphere, waves from this scale range clearly play
454 a dominant role, more than 90% over broad latitude regions in both hemispheres, and
455 up to 60% over the equator. This dominance underscores the essential need for gravity
456 wave parameterization in the stratosphere, mesosphere and lower thermosphere in coarse
457 resolution models. The smaller scale contribution can still be large in the thermosphere,
458 up to 80% at higher latitudes. It is noted that the percentage contribution by the smaller
459 scale waves can be overestimated by this method, when the momentum flux spectrum
460 is over flattened when the zonal wind become too large due to insufficient gravity wave
461 forcing in the model (H.-L. Liu, 2019).

462 **3.4 Neutral and plasma perturbations caused by gravity waves**

463 As discussed in previous sections, gravity waves cause perturbations in neutral winds
464 and temperature in the thermosphere, and the perturbations show strong local time de-
465 pendence. The wave structure in zonal wind and its dependence on local time can be
466 clearly seen in Figure 13, with the zonal wind perturbations tilt eastward into the west-
467 ward wind (local daytime) and westward into the eastward wind (local nighttime). This
468 is most evident above ~ 120 km and at all latitudes. Modulation of gravity waves by the
469 migrating semi-diurnal tide (SW2) in the lower thermosphere (below ~ 120 km) is seen
470 at 50°N where SW2 is strong.

471 To better understand the wave effects in the thermosphere and ionosphere, we ex-
472 amine most closely the neutral and plasma perturbations near the F-region peak. A high-
473 pass filter using the Savitzky-Golay method (Savitzky & Golay, 1964) is applied to neu-
474 tral meridional wind, electron density, total electron content, and the zonal and verti-
475 cal components of $\mathbf{E} \times \mathbf{B}$ drifts. Two period ranges are examined: shorter than 2 hours
476 (0.14 mHz), and shorter than 20 minutes (0.83 mHz).

477 Figure 14(left) shows the neutral meridional wind perturbation in the F-region with
 478 period shorter than 2 hours. Clear wave signature is seen, with the largest amplitude (up
 479 to 80 ms^{-1}) found at mid to high latitudes in the winter hemisphere. The wave prop-
 480 agation direction has an evident local time dependence, with the propagation direction
 481 being generally opposite to the large-scale F-region wind. For example, the wind changes
 482 quickly from northeastward on the night side to northwestward on the day side around
 483 the terminator at mid-high northern latitudes. As such, the propagation direction of the
 484 gravity waves changes from southwestward on the night side to southeastward on the day
 485 side. Meridional wind perturbations with periods shorter than 20 min can still be quite
 486 strong at mid to high winter latitudes, up to 25 ms^{-1} (Figure 14, right). Their spatial
 487 distribution is similar to those shown in left figure, and the spatial scales are generally
 488 smaller.

489 Figure 15 (left panel) are the electron density perturbations (relative to the unfilter-
 490 ed value) near the F-region peak with periods less than 2 hours and less than 20 min-
 491 utes. The spatial distribution and local time dependence of the perturbations and their
 492 propagation are very similar to gravity wave signatures seen in the meridional wind (Fig-
 493 ure 14), suggesting the gravity waves as the main driver of these traveling ionospheric
 494 disturbances (TIDs). The relative perturbation amplitude associated with the waves is
 495 about $\pm 15\%$ (< 2 hours) and $\pm 5\%$ (< 20 min). Similar features are also seen in the to-
 496 tal electron content (TEC) (Figure 15 right panel), with magnitudes of $\sim \pm 3\%$ (< 2 hours)
 497 and $\sim \pm 1\%$ (< 20 min). It is also worth noting that the consistent southwestward prop-
 498 agation of neutral wind perturbations and TIDs at night time suggests that gravity waves
 499 can provide an alternative mechanism to Perkins instability in driving this propagation
 500 pattern. The large relative changes around the terminators are artifacts from the filter-
 501 ing. An animation of the relative TEC perturbations (period less than 2 hours) from one
 502 day of simulation (January 13) has been included as Supporting Information (Movie S1).

503 Both wind and electron density perturbations are significant according to these sim-
 504 ulations. The field aligned drift perturbation associated with the neutral wind would be
 505 of similar magnitude. It is noteworthy that the frequency range overlaps with that of
 506 ultra-low frequency (ULF) waves (Pc5-6, period shorter than 20 minutes). These grav-
 507 ity wave driven perturbations may thus play a role in magnetosphere, ionosphere and
 508 thermosphere coupling. This will be explored in the future using coupled geospace mod-
 509 els, such as the Multiscale Atmosphere-Geospace Environment (MAGE) model ([https://](https://cgs.jhuapl.edu/)
 510 cgs.jhuapl.edu/).

511 Wave oscillations of wind and electron density (thus electric conductivities) per-
 512 turb the $\mathbf{E} \times \mathbf{B}$ drift (Figure 16). Both the zonal and vertical components of $\mathbf{E} \times \mathbf{B}$ have
 513 the largest amplitudes around the geomagnetic equator, up to $\sim 5 \text{ ms}^{-1}$ (< 2 hours), 1.5
 514 ms^{-1} (zonal, < 20 min) and 0.8 ms^{-1} (vertical, < 20 min). The strongest perturbations
 515 are found during night time, probably because the high frequency wind perturbations
 516 are generally stronger in the F-region, and the F-region dynamo is dominant during night-
 517 time. While the phase lines of the perturbations are perpendicular to the geomagnetic
 518 equator, they curve toward eastward or westward directions away from the equator, with
 519 a "C" or reverse "C" shape. This likely reflects the relative position of the wave phase
 520 lines of the wind perturbations with respect to the field lines. A more quantitative study
 521 will be conducted in the future.

522 4 Summary and Conclusions

523 With the newly developed species-dependent spectral element dynamical core, WACCM-
 524 X simulations can now be performed with high spatial resolution: ~ 25 km horizontally
 525 and 0.1 scale height vertically. The high-resolution simulation for January has been an-
 526 alyzed to study the gravity wave distribution and the wave effects in the thermosphere
 527 and ionosphere. The overall wave activity continues to increase with altitude in the ther-

528 mosphere, with horizontal wind and temperature perturbations maximizing between 110–
 529 200 km, and vertical wind between 200–300 km. The largest wave activities are located
 530 at higher latitudes, with larger magnitude in the northern (winter) hemisphere. In the
 531 winter hemisphere, the longitudinal distribution of gravity waves from the mesosphere
 532 to the upper thermosphere is similar to that in the stratosphere, with the strongest ac-
 533 tivities extending from eastern North America, across the North Atlantic and to west-
 534 ern Eurasia. The winter stratospheric jet may play an important role in exciting these
 535 waves. In the summer hemisphere, the peak wave activities shift from tropical latitudes
 536 at lower altitudes to higher latitudes with increasing height up to the upper mesosphere,
 537 coincident with the altitudes with zonal wind reversal and/or the large eastward shear
 538 of the mean zonal wind. Gravity waves display strong local time dependence in the up-
 539 per atmosphere due to the modulation by tidal waves and the large day-night change
 540 of large-scale atmosphere state. In particular, the thermospheric gravity waves propa-
 541 gate predominantly against the large-scale wind. These waves have larger intrinsic fre-
 542 quencies, vertical wavelengths and propagation speeds due to doppler shift, and are thus
 543 less vulnerable to the strong molecular damping. The strongest wave activities in the
 544 upper thermosphere are near magnetic poles in both hemisphere. This could be due to
 545 reduced ion drag for gravity waves with large vertical wavelengths, whose phase lines be-
 546 come more parallel with the fieldlines. The features of wave distribution are similar to
 547 available observations in the middle and upper atmosphere.

548 The dissipation of the gravity waves in the thermosphere leads to strong forcing
 549 in the zonal and meridional directions. In January, the mean zonal/meridional forcing
 550 is found to be predominantly eastward/southward in both hemispheres, with the largest
 551 values found at mid-high latitudes between 200–300 km (exceeding $100 \text{ ms}^{-1}\text{d}^{-1}$ in the
 552 NH). The Coriolis forcing associated with the eastward gravity wave forcing is equator-
 553 ward. Therefore the zonal and meridional forcings both tend to slow down the mean merid-
 554 ional circulation in the NH, while in the SH they offset each other. The thermospheric
 555 circulation changes the compositional structure: the slowdown of meridional/downward
 556 circulation in the winter hemisphere reduces the column integrated O/N_2 . This ratio is
 557 further reduced by wave induced transport. These changes result in improved agreement
 558 with observations. The gravity wave forcing, the mean circulation and the tides change
 559 significantly from day to day.

560 The zonal spectra of the kinetic energy and the vertical momentum flux still fol-
 561 low power-law distributions in the thermosphere. The spectral slopes of the latter is gen-
 562 erally shallower than the former, as in the lower and middle atmosphere. The power-law
 563 distributions allow the examination of the scale dependence of kinetic energy and mo-
 564 mentum flux. For kinetic energy, larger scale waves are dominant at most latitudes. For
 565 momentum flux, on the other hand, smaller scale waves can contribute significantly, es-
 566 pecially at middle to high latitudes.

567 Gravity waves are an important driver of traveling atmosphere disturbances (TADs)
 568 and traveling ionosphere disturbances (TIDs). The waves and these disturbances depend
 569 sensitively on the background wind in the thermosphere, which is in turn local time de-
 570 pendent. This is because the molecular damping is inversely proportional to the square
 571 of the vertical wavelength, which changes as a result of Doppler shift by the background
 572 wind. The magnitude of the wave induced TID in TEC is $\sim 3\%$ (periods < 2 hours), with
 573 temporal and spatial structures similar to the F-region electron density perturbations
 574 and the gravity waves at F-region altitudes. Even with this hydrostatic model, the sim-
 575 ulated thermospheric and ionospheric perturbations still have significant power at short
 576 periods (less than 20 min). Its role in magnetosphere-ionosphere-thermosphere coupling
 577 is worth further investigation in future studies.

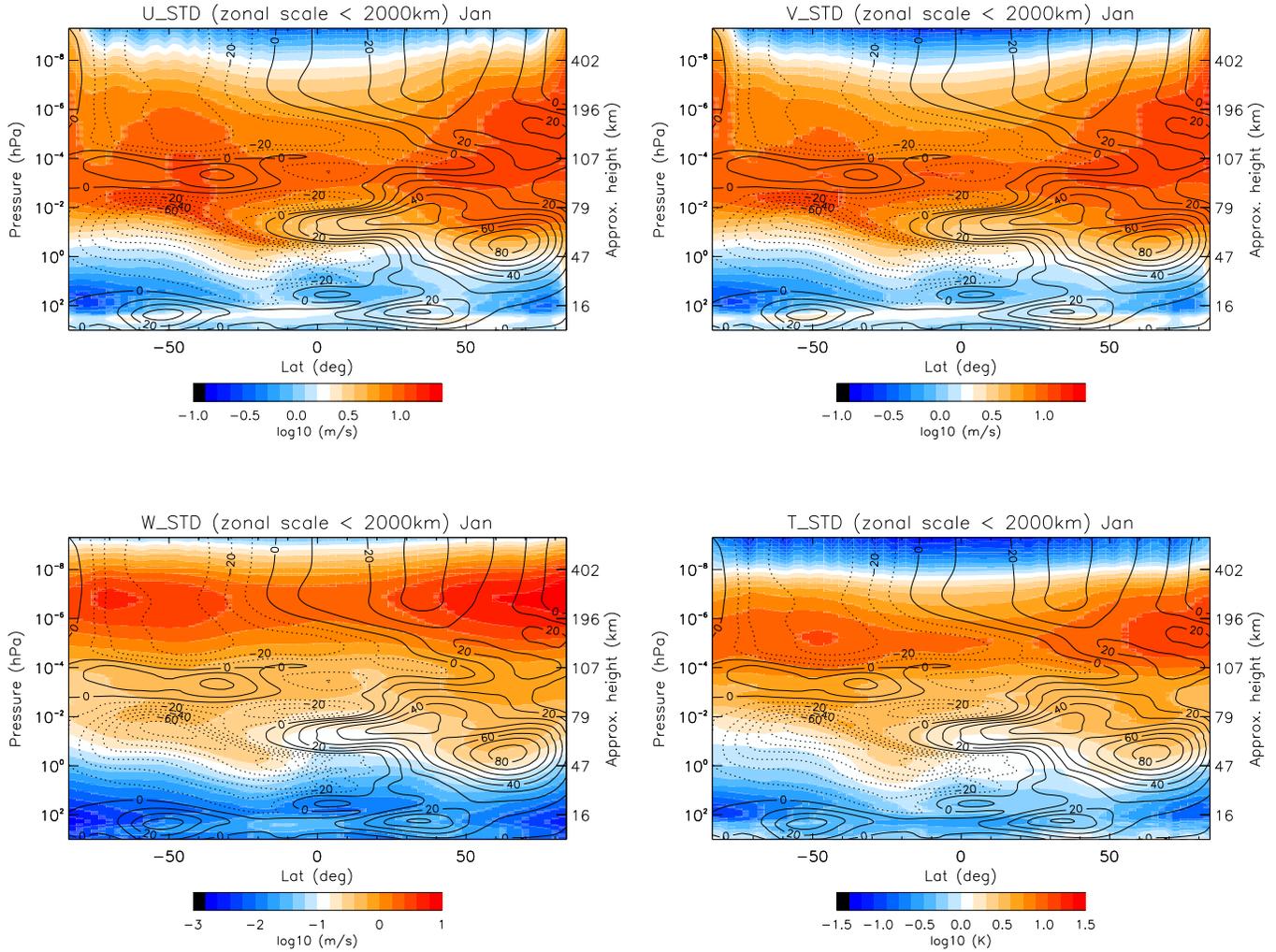


Figure 1: Zonally averaged standard deviation of (upper left) zonal, (upper right) meridional, (lower left) vertical winds and (lower right) neutral temperature perturbations with zonal scales less than 2000 km for January. Contour lines are zonal mean zonal wind (Solid: eastward, with contour interval of 10 ms^{-1}).

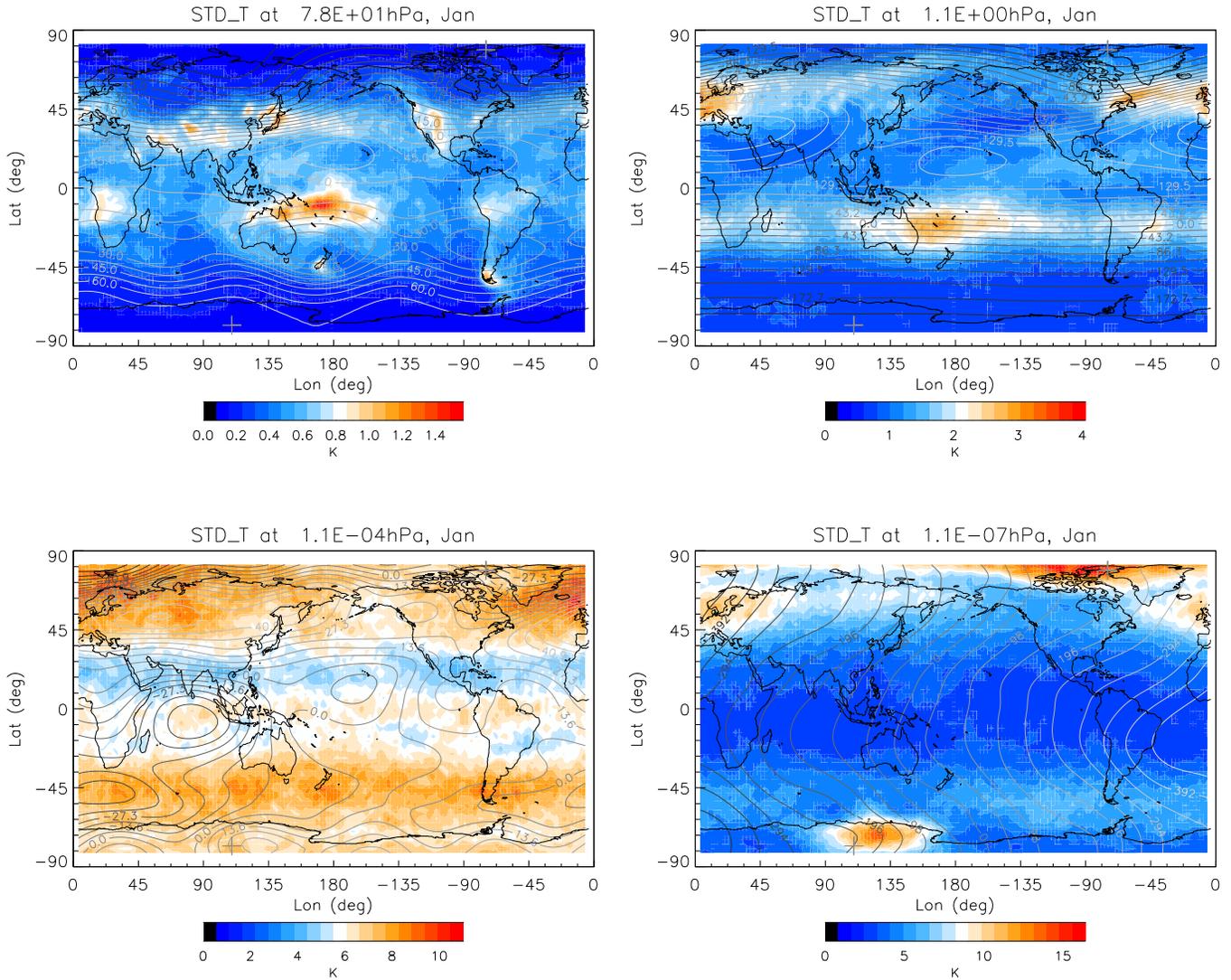


Figure 2: Standard deviation of temperature perturbations at (upper left) 78, (upper right) 1.1, (lower left) 1.1×10^{-4} and (lower right) 1.1×10^{-7} hPa, averaged over four UT times (0, 6, 12, and 18 hours) for January. The contour lines are stream functions calculated from the horizontal winds. Contour lines with lighter shades have larger values. Atmosphere flow is toward the right of the down-gradient direction.

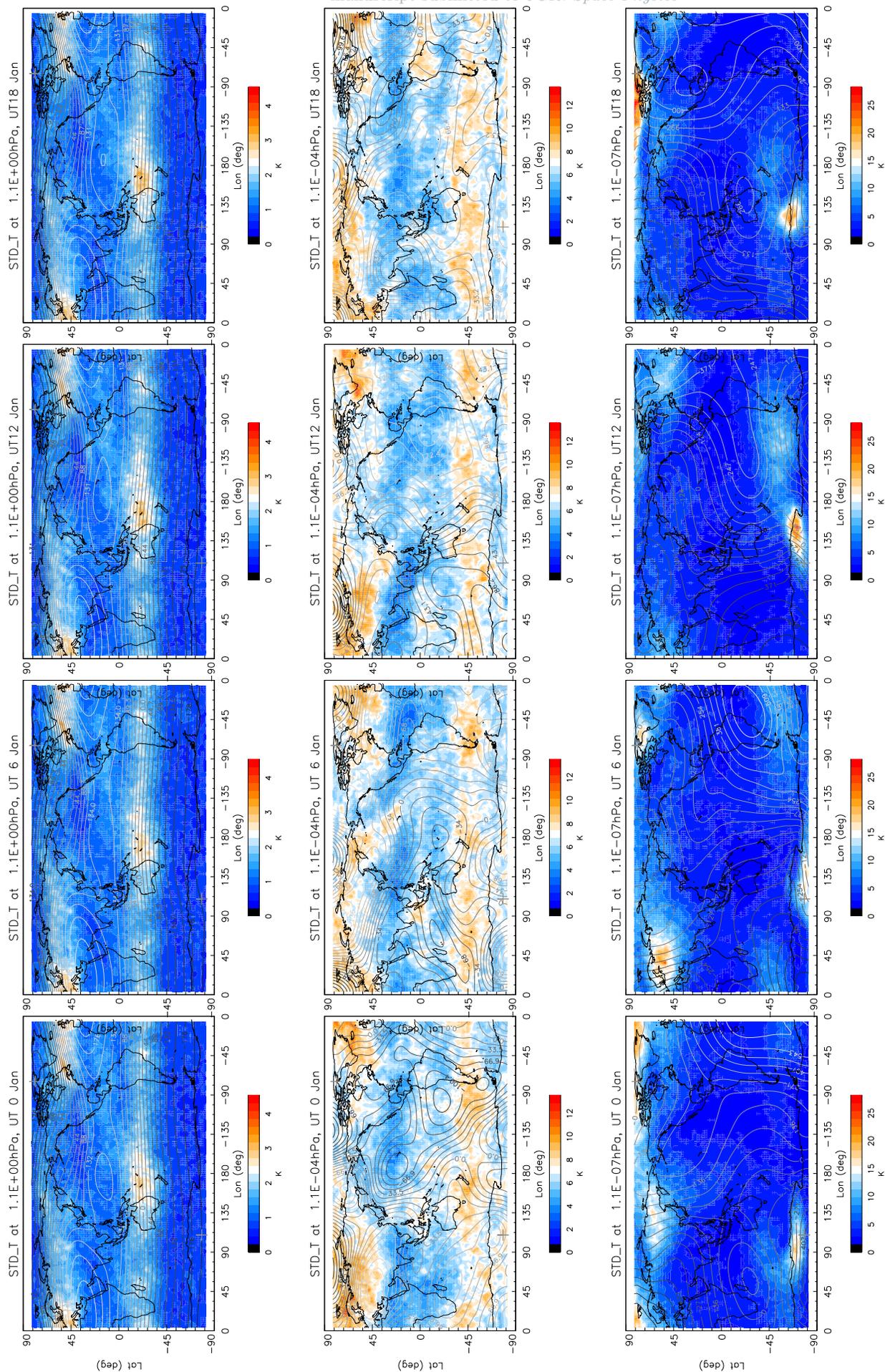


Figure 3: Standard deviation of temperature perturbations at (upper panels) 1.1×10^{-4} and (lower panels) 1.1×10^{-7} hPa at four UT (0, 6, 12, and 18 hours, from left to right) for January. The contour lines are stream functions calculated from the horizontal winds. Contour lines with lighter shades have larger values. Atmosphere flow is toward the right of the down-gradient direction.

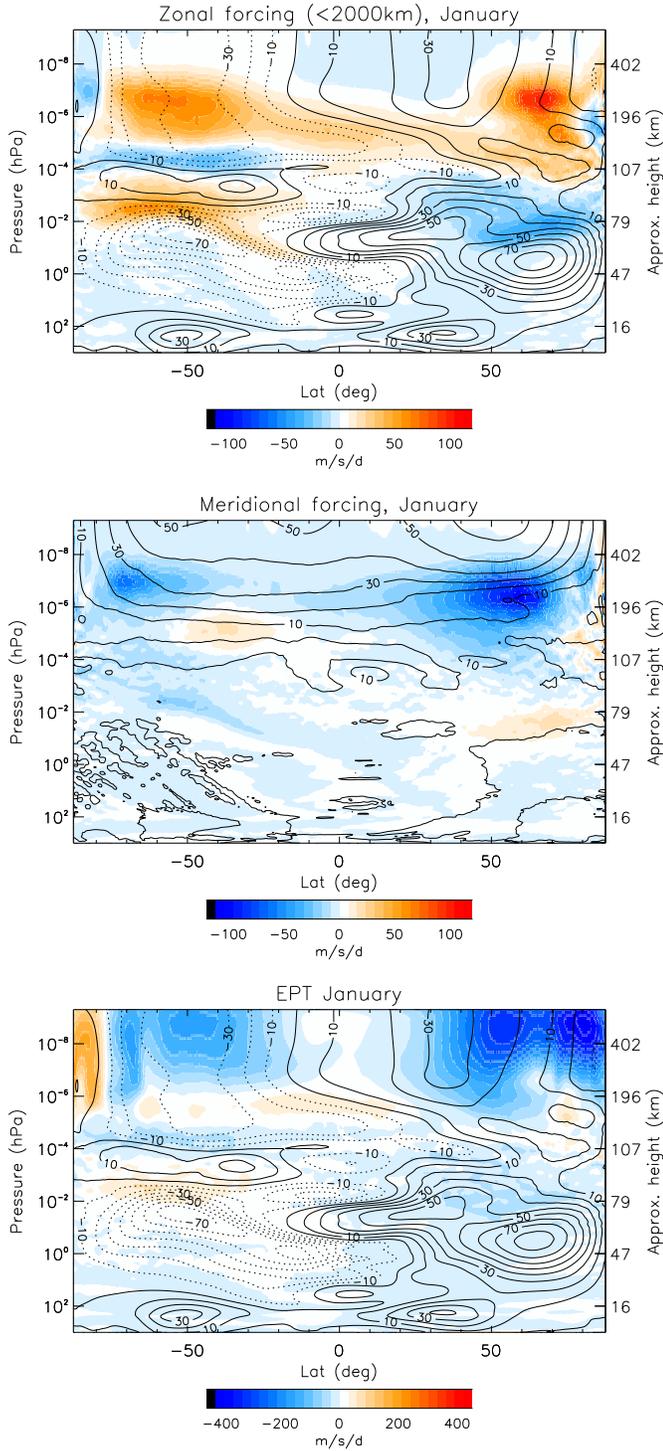


Figure 4: Zonal mean (upper panel) zonal forcing and (middle panel) meridional forcing by gravity waves with zonal scales less than 2000 km. Contour lines are zonal mean zonal wind in the upper panel (solid: eastward) and zonal mean meridional wind in the middle panel (solid: northward). Contour intervals: 10ms^{-1} . Lower panel: Total zonal mean zonal forcing by all resolved waves. Contour lines are the same as in the upper panel.

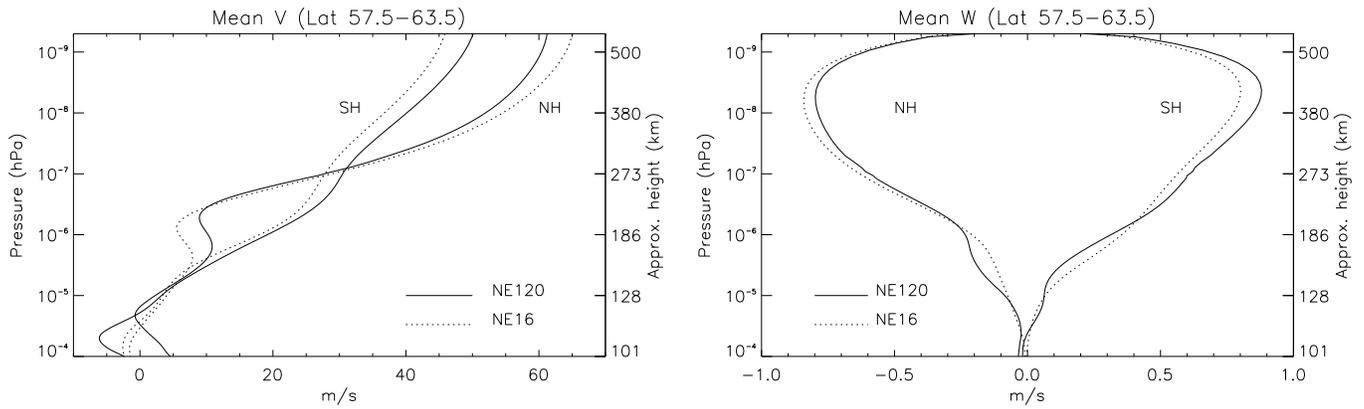


Figure 5: Zonal mean (left) meridional and (right) vertical winds in the thermosphere averaged over $57.5\text{--}63.5^\circ$ latitudes in both hemispheres. The solid lines are from the high resolution simulation, while the dotted lines are from the coarse resolution simulation.

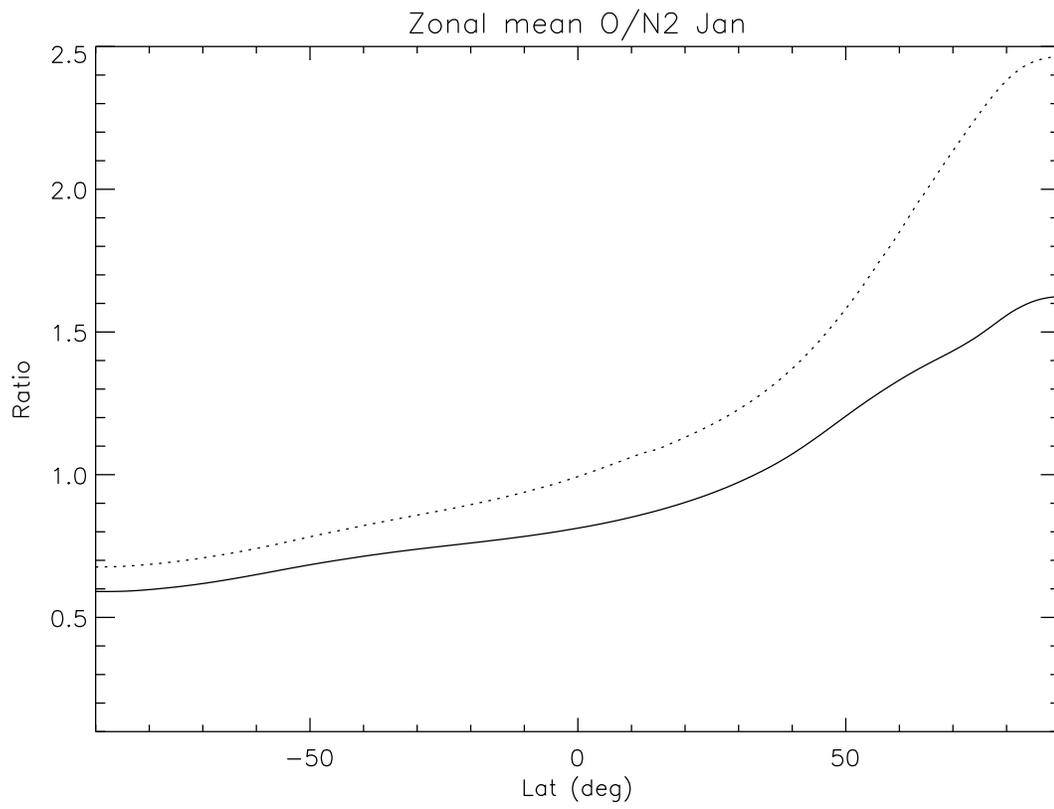


Figure 6: Zonally averaged O/N₂ (column integrated) from high resolution (solid) and coarse resolution (dotted) WACCM-X simulations.

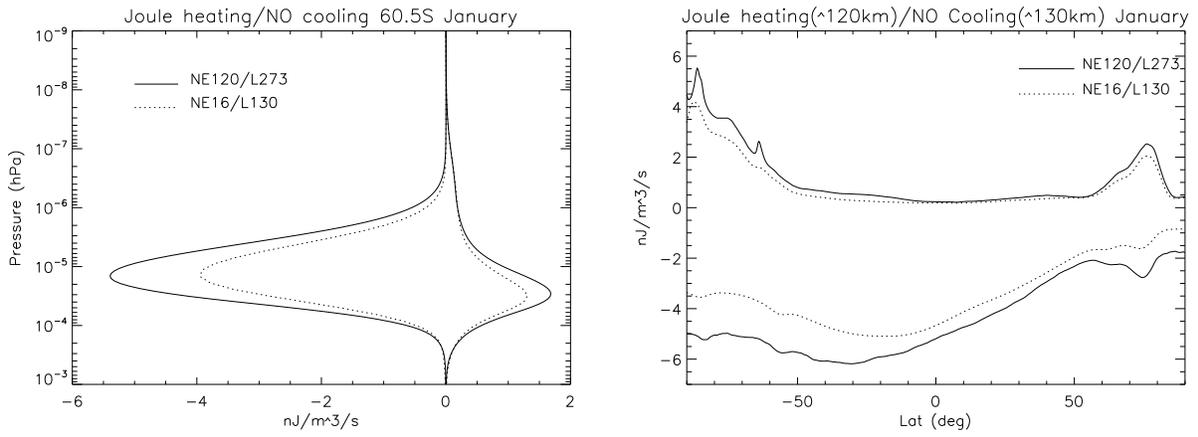


Figure 7: Vertical (left) and latitudinal (right) profiles of Joule heating and NO cooling from high-resolution (solid lines) and coarse-resolution (dotted lines) simulations.

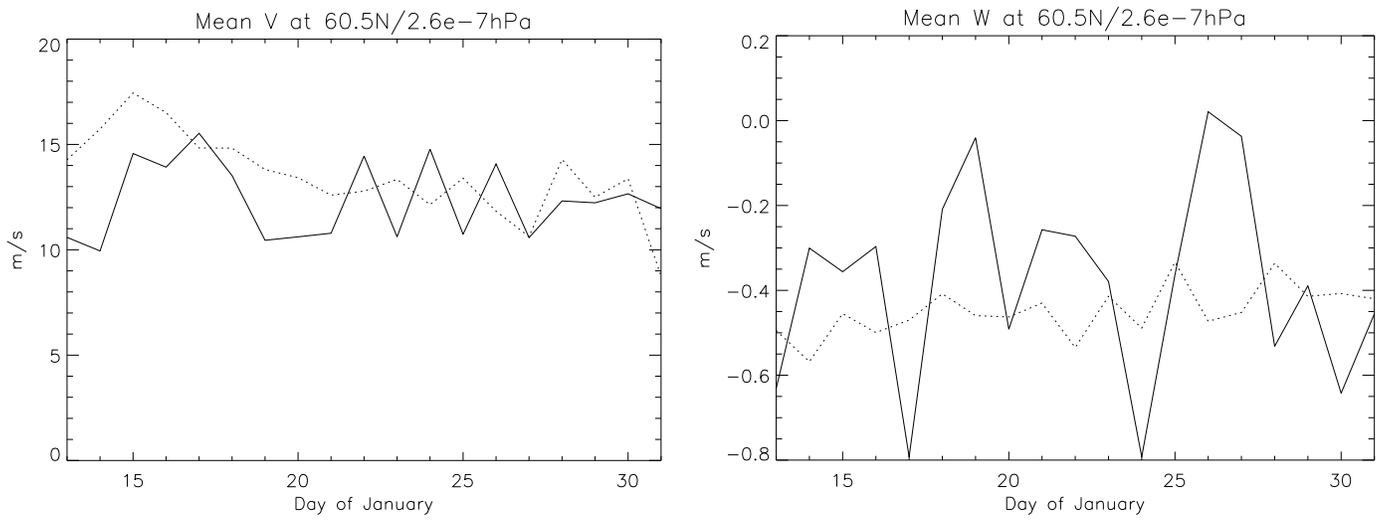


Figure 8: Zonal mean (left) meridional and (right) vertical winds at 60.5°N and 2.6×10^{-7} hPa for 13–31 January. Solid lines: high resolution simulation. Dotted lines: coarse resolution simulation.

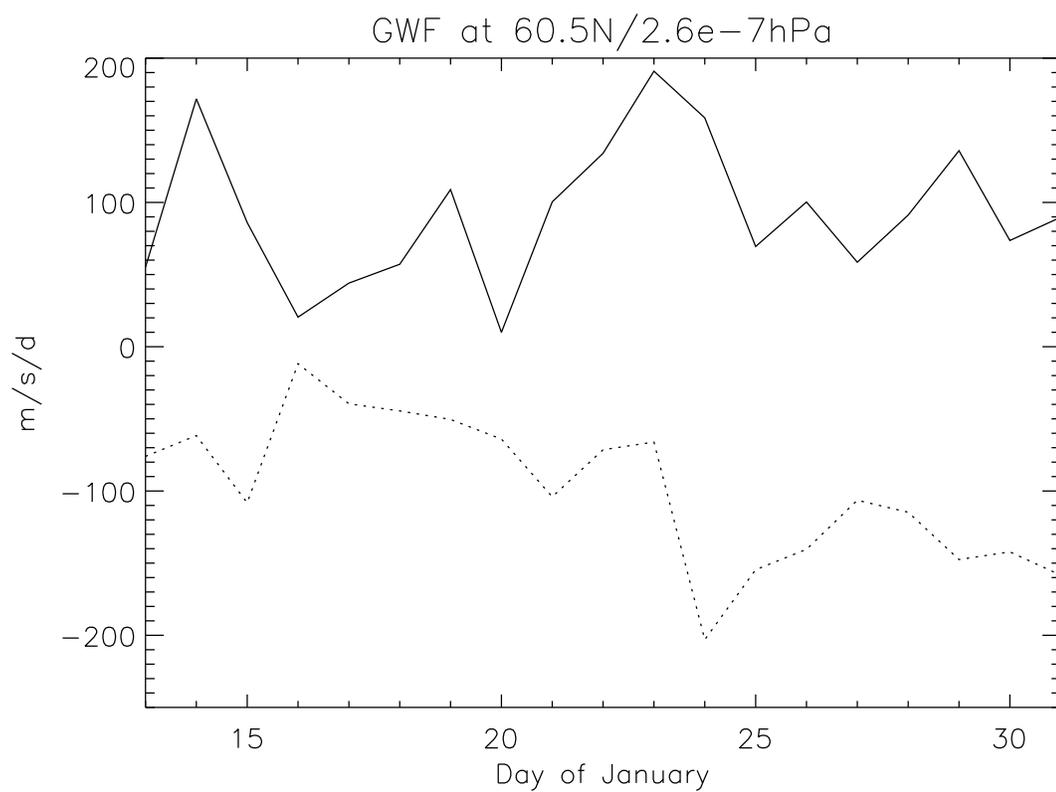


Figure 9: Zonal mean zonal (solid line) and meridional (dotted line) forcing at 60.5°N latitude and 2.6×10^{-7} hPa for 13–31 January.

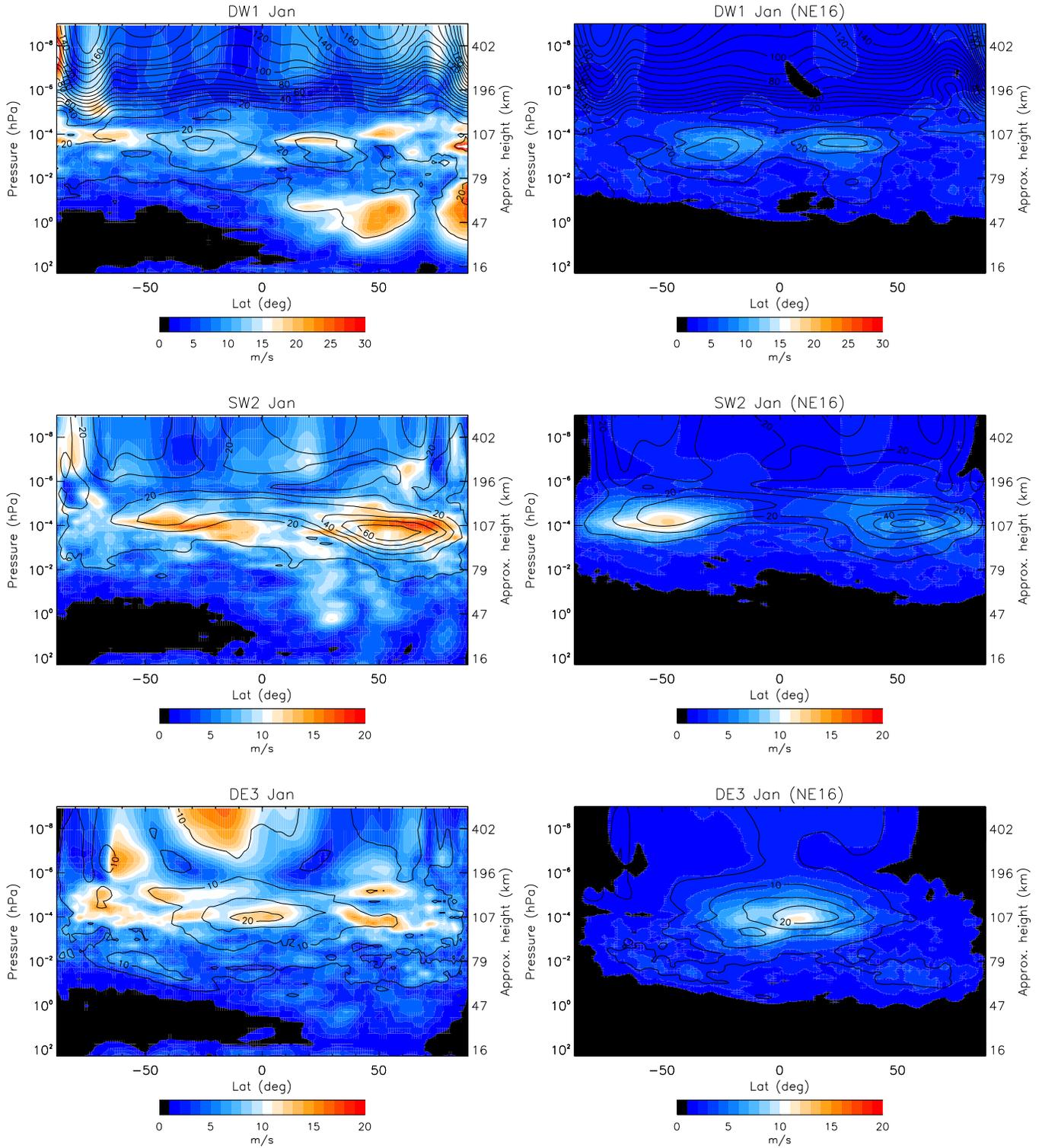


Figure 10: The mean (line contour) and standard deviation (color contour) of the DW1 (upper), SW2 (middle) and DE3 (lower) tidal components from the NE120 (left) and NE16 (right) simulations. Line contour intervals: 10ms^{-1} (upper and middle) and 5ms^{-1} (lower).

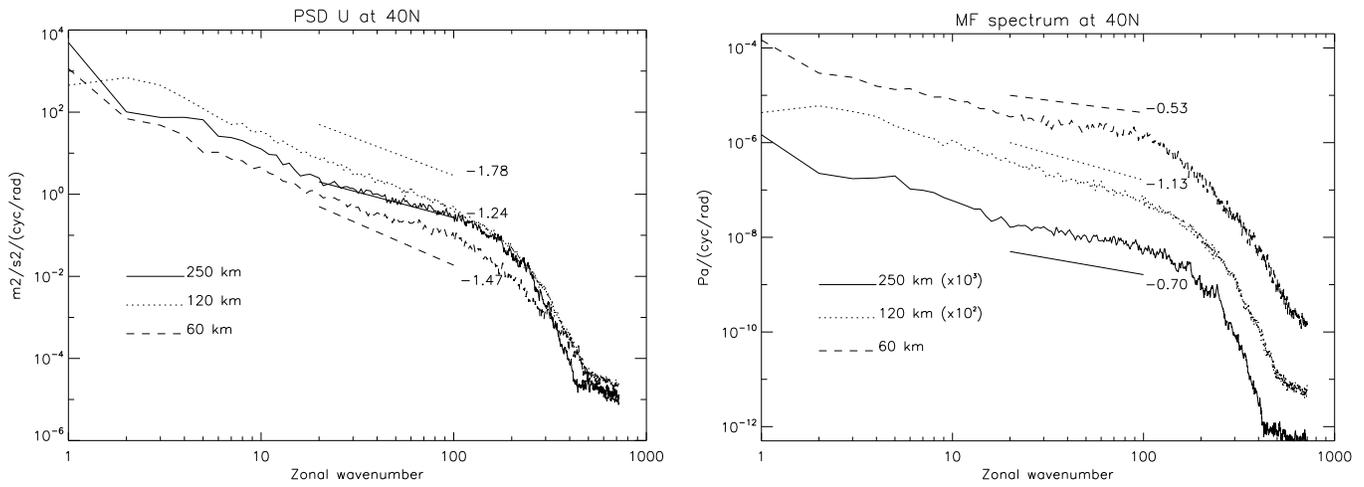


Figure 11: Zonal wind kinetic energy spectra (left) and momentum flux spectra (right) at 40°N and 3 different altitudes (60, 120 and 250 km). Spectral slopes for the mesoscale range (zonal wavenumber 20–100) are shown by the straight lines.

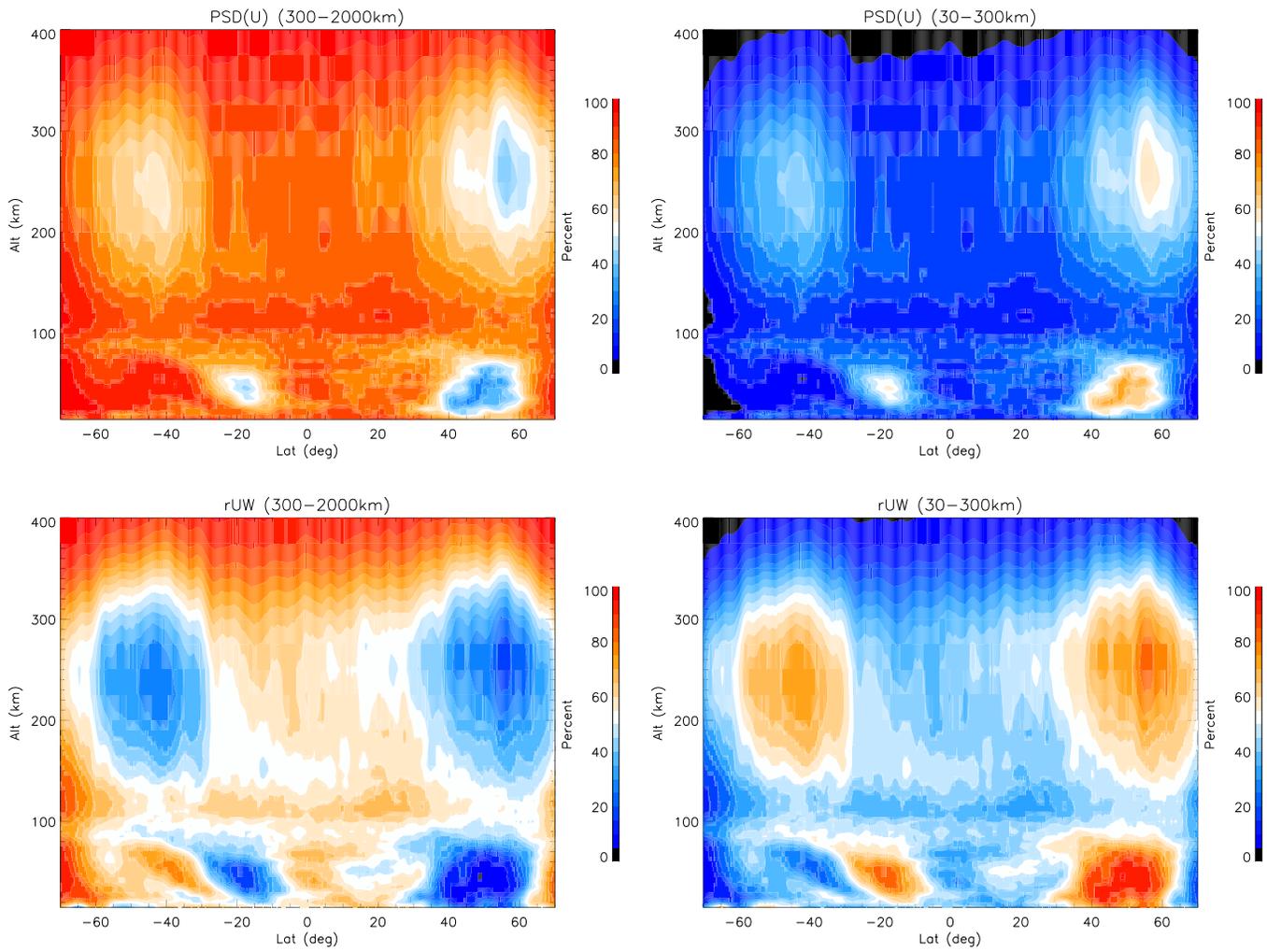


Figure 12: Percentage contributions to the total (upper panels) zonal kinetic energy and (lower panels) vertical flux of zonal momentum by (left panels) larger zonal scale waves and (right panels) smaller zonal scale waves.

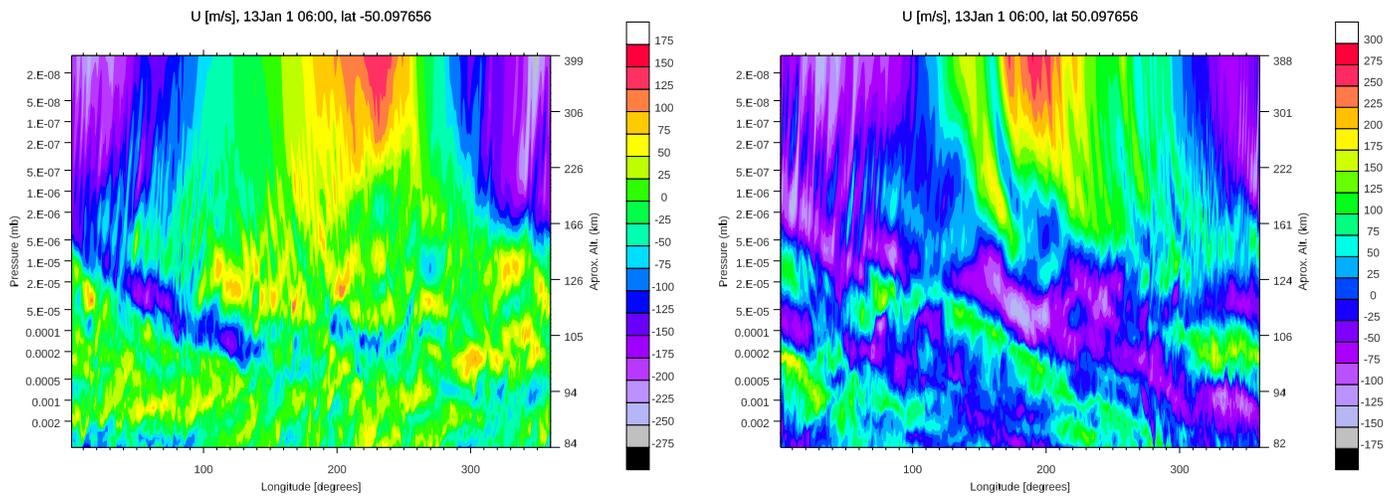


Figure 13: Upper mesosphere and thermosphere zonal wind at (left) 50°S and (right) 50°N and UT 6 hour on January 13.

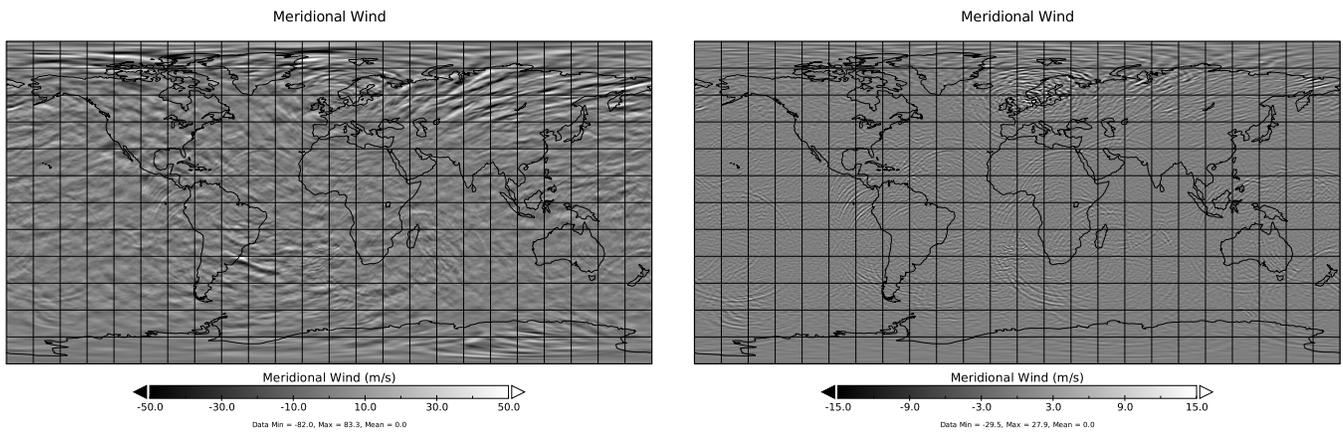


Figure 14: High-pass filtered meridional wind, with period shorter than (left panel) 2 hours and (right panel) 20 minutes. The UT time is 3 hour on January 30th.

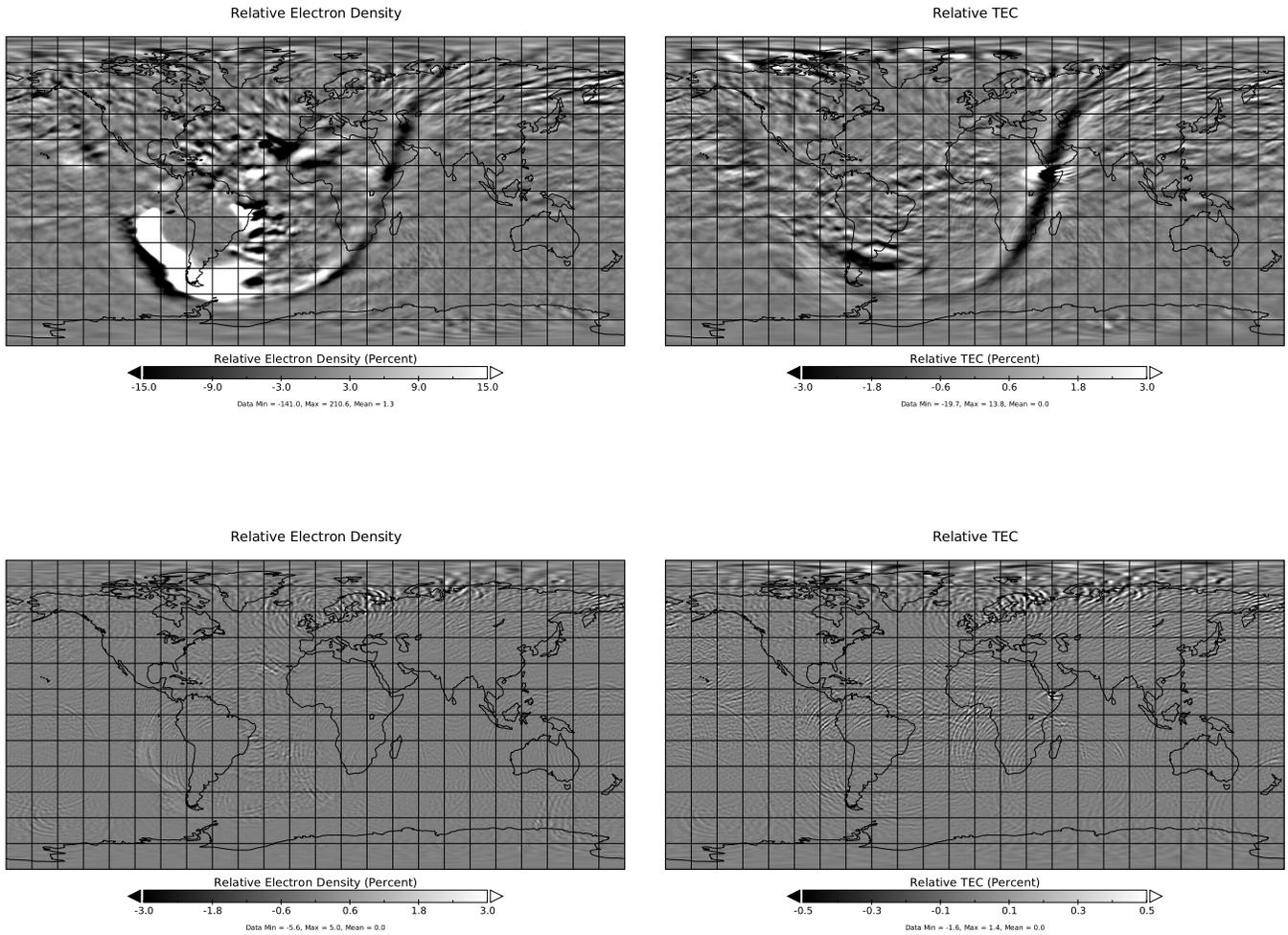


Figure 15: High-pass filtered (left panels) relative electron density perturbations near F region peak and (right panels) relative total electron content (TEC) perturbations, with period shorter than (upper panels) 2 hours and (lower panels) 20 minutes. The UT time is 3 hour on January 30th.

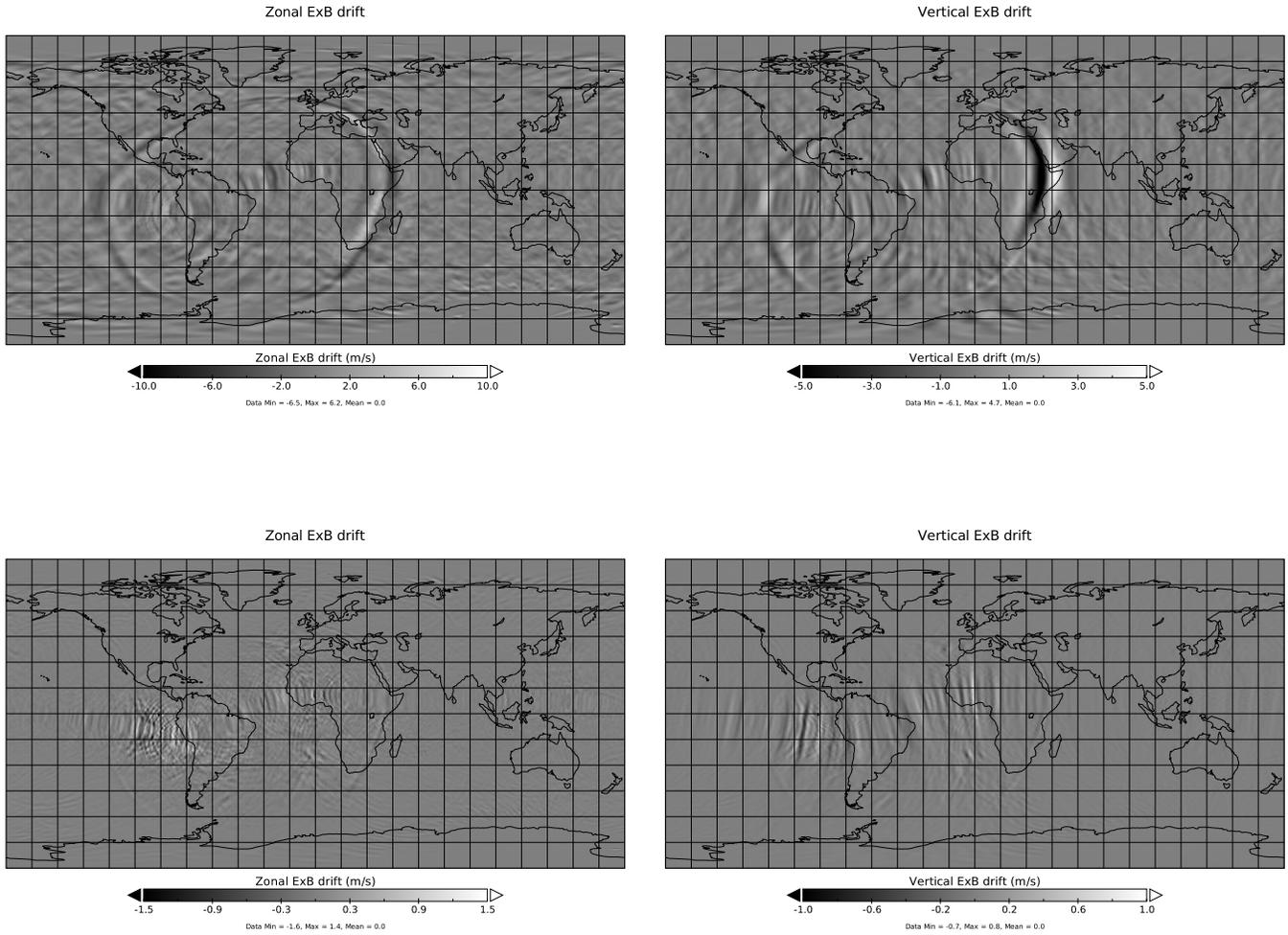


Figure 16: High-pass filtered (left panels) zonal and (right panels) vertical components of $E \times B$ drifts, with period shorter than (upper panels) 2 hours and (lower panels) 20 minutes. The UT time is 3 hour on January 30th.

578

Appendix A Species dependent thermodynamics

Lauritzen et al. (2018) derived energy consistent equations of motion and equation of state for moist air including water vapor and condensates. Here we expand that framework for species dependent dry air. The implementation is such that the user can specify a list of major species via the namelist. Let \mathcal{L}_{dry} be the set of species that make up dry air that in this study is given by

$$\mathcal{L}_{dry} = \{O, O_2, H, N_2\}. \quad (A1)$$

The framework can easily be extended to include more species in dry air. The set of all components of moist air is given by

$$\mathcal{L}_{all} = \mathcal{L}_{dry} \cup \mathcal{L}_{H_2O}, \quad (A2)$$

where \mathcal{L}_{H_2O} is the set of water species (water vapor, cloud liquid, ice, rain, and snow):

$$\mathcal{L}_{H_2O} = \{wv', cl', ci', rn', sn'\}. \quad (A3)$$

The dry mixing ratio for each species is defined as

$$m^{(\ell)} \equiv \frac{\rho^{(\ell)}}{\rho^{(d)}}, \quad (A4)$$

where $\rho^{(d)}$ is the mass of dry air per unit volume of moist air and $\rho^{(\ell)}$ is the mass of the species ℓ per unit volume of moist air. Note that the mixing ratio for dry air is one by definition

$$\sum_{\ell \in \mathcal{L}_{dry}} m^{(\ell)} = 1, \quad (A5)$$

and that the mixing ratio for N_2 is derived from the other components of dry air

$$m^{(N_2)} = 1 - m^{(O)} - m^{(O_2)} - m^{(H)}. \quad (A6)$$

The specific (moist) mixing ratios are given by

$$q^{(\ell)} \equiv \frac{\rho^{(\ell)}}{\rho}, \quad (A7)$$

579

where ρ is the density of moist air.

580

A1 Species dependent specific heat for dry air

The specific heat at constant pressure for species ℓ is given by

$$c_p^{(\ell)} = \frac{1}{2} R_d \frac{dof^{(\ell)}}{\mathcal{M}^{(\ell)}}, \quad (A8)$$

where R_d is the universal gas constant, $\mathcal{M}^{(\ell)}$ is the molecular mass of species ℓ and $dof^{(\ell)}$ is degrees of freedom:

$$dof = \{5, 7, 7, 5\} \text{ for set } \mathcal{L}_{dry}. \quad (A9)$$

581

The formula for the specific heat of dry air at constant pressure is simply the sum of the specific heats for each major species

582

$$c_p^{(d)} = \sum_{\ell \in \mathcal{L}_{dry}} c_p^{(\ell)} m^{(\ell)}. \quad (A10)$$

583

The generalized specific heat for species dependent moist air is then

$$c_p = \frac{c_p^{(d)} + \sum_{\ell \in \mathcal{L}_{H_2O}} c_p^{(\ell)} m^{(\ell)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}}, \quad (A11)$$

$$= \sum_{\ell \in \mathcal{L}_{all}} c_p^{(\ell)} q^{(\ell)}, \quad (A12)$$

584

(Lauritzen et al., 2018). There is currently a discrepancy between CAM physics (that assumes that $\mathcal{L}_{H_2O} = \{wv'\}$) and the spectral-element dynamical core that includes all forms of water in the thermodynamics.

585

586

587

A2 Species dependent R for dry air

588

The definition of the gas constant for dry air, according to the original definition from kinetic theory, is

589

$$R^{(d)} = R_d \sum_{\ell \in \mathcal{L}_{dry}} \left[\frac{m^{(\ell)}}{\mathcal{M}^{(\ell)}} \right], \quad (\text{A13})$$

$$= \sum_{\ell \in \mathcal{L}_{dry}} \frac{R_d}{\mathcal{M}^{(\ell)}} m^{(\ell)}, \quad (\text{A14})$$

$$= \sum_{\ell \in \mathcal{L}_{dry}} R^{(\ell)} m^{(\ell)} \quad (\text{A15})$$

where $R^{(\ell)}$ is given by

$$R^{(\ell)} = \frac{R_d}{\mathcal{M}^{(\ell)}}, \text{ for } \ell \in \mathcal{L}_{dry}, \quad (\text{A16})$$

590

in which case the generalized R becomes

$$R = \frac{R^{(d)} + \sum_{\ell \in \mathcal{L}_{H_2O}} R^{(\ell)} m^{(\ell)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}}, \quad (\text{A17})$$

$$= \sum_{\ell \in \mathcal{L}_{all}} \left[R^{(\ell)} q^{(\ell)} \right], \quad (\text{A18})$$

591

where $R^{(\ell)} = 0$ for non-gas components of air (condensates).

592

A3 Virtual temperature

Let \mathcal{L}_{gas} be the set of gaseous components of dry air. Each gaseous component of air satisfies in the ideal gas law

$$p^{(\ell)} V^{(gas)} = V \rho^{(\ell)} R^{(\ell)} T, \text{ for } \ell \in \mathcal{L}_{gas} \quad (\text{A19})$$

593

where $p^{(\ell)}$ is the partial pressure of gas ℓ , V is the volume of moist air and $V^{(gas)}$ the volume of the gaseous components of moist air. Applying Dalton's law of partial pressures, the total pressure is given by

594

595

$$\begin{aligned} p &= \sum_{\ell \in \mathcal{L}_{gas}} p^{(\ell)}, \\ &= \frac{V}{V^{(gas)}} \sum_{\ell \in \mathcal{L}_{gas}} \left[\rho^{(\ell)} R^{(\ell)} T \right], \text{ using (A19)} \\ &= \frac{V}{V^{(gas)}} \rho^{(d)} R^{(d)} T \sum_{\ell \in \mathcal{L}_{gas}} \left[\frac{\rho^{(\ell)} R^{(\ell)}}{\rho^{(d)} R^{(d)}} \right], \text{ 'pull' out } \rho^{(d)} R^{(d)}, \\ &\approx \rho^{(d)} R^{(d)} T \sum_{\ell \in \mathcal{L}_{gas}} \left[\frac{\rho^{(\ell)} R^{(\ell)}}{\rho^{(d)} R^{(d)}} \right], \text{ assume condensates occupy 0 volume } \frac{V}{V^{(gas)}} = 1, \\ &= \rho^{(d)} R^{(d)} T \sum_{\ell \in \mathcal{L}_{gas}} \left[m^{(\ell)} \frac{R^{(\ell)}}{R^{(d)}} \right], \text{ use } m^{(\ell)} \equiv \frac{\rho^{(\ell)}}{\rho^{(d)}}, \\ &= \rho R^{(d)} T \sum_{\ell \in \mathcal{L}_{gas}} \left[\frac{m^{(\ell)} \frac{R^{(\ell)}}{R^{(d)}}}{\sum_{\mathcal{L}_{all}} m^{(\ell)}} \right], \text{ using } \rho = \rho^{(d)} \sum_{\mathcal{L}_{all}} m^{(\ell)}, \\ &= \rho R^{(d)} T \sum_{\ell \in \mathcal{L}_{all}} \left[\frac{m^{(\ell)} \frac{R^{(\ell)}}{R^{(d)}}}{\sum_{\mathcal{L}_{all}} m^{(\ell)}} \right], \text{ since } R^{(\ell)} = 0 \text{ for non-gas components,} \\ &= \rho R^{(d)} T \frac{R}{R^{(d)}}, \text{ where } R^{(d)} \text{ is given by (A17),} \\ &= \rho R^{(d)} T_v, \end{aligned}$$

596 where the virtual temperature is given by

$$T_v = T \frac{R}{R^{(d)}}, \quad (\text{A20})$$

$$= T \left[\frac{1 + \left(\frac{R^{(wv)}}{R^{(d)}} \right) m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} \right], \quad (\text{A21})$$

597 and $R^{(d)}$ is given in (A17). We can rewrite (A21) as

$$T_v = T \left[\frac{1 + \left(\frac{R^{(wv)}}{R^{(d)}} \right) m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} \right] \quad (\text{A22})$$

$$= T \left[\frac{1 + m^{(wv)} + \left(\frac{R^{(wv)}}{R^{(d)}} \right) m^{(wv)} - m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} \right] \quad (\text{A23})$$

$$= T \left[\frac{1 + m^{(wv)} + \left\{ \left(\frac{R^{(wv)}}{R^{(d)}} \right) - 1 \right\} m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} \right] \quad (\text{A24})$$

$$= T \left[\frac{1 + m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} + \frac{\left\{ \left(\frac{R^{(wv)}}{R^{(d)}} \right) - 1 \right\} m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} \right] \quad (\text{A25})$$

$$= T \left[\frac{1 + m^{(wv)}}{\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)}} + \left\{ \left(\frac{R^{(wv)}}{R^{(d)}} \right) - 1 \right\} q^{(wv)} \right] \quad (\text{A26})$$

So if and only if water vapor is the only non-dry component of dry air, i.e.

$$\sum_{\ell \in \mathcal{L}_{all}} m^{(\ell)} = 1 + m^{(wv)}, \quad (\text{A27})$$

then (A21) can be written as

$$T_v = T \left[1 + \left(\frac{R^{(wv)}}{R^{(d)}} - 1 \right) q^{(wv)} \right], \quad (\text{A28})$$

598 which is the formula used in CAM physics but not the spectral-element dynamical core.

599 Appendix B Molecular viscosity and thermal conductivity

The implementation of molecular viscosity and thermal conductivity in WACCM-X is split into a horizontal part (handled by the dynamical core) and vertical (performed in physics). That is, for the molecular viscosity the dynamical core solves

$$\frac{d\vec{v}}{dt} = \frac{1}{\rho} \nabla_z \cdot (kmvis \nabla_z \vec{v}), \quad (\text{B1})$$

where $kmvis$ is a dynamic viscosity coefficient which is a function of the major species, and physics solves

$$\frac{d\vec{v}}{dt} = \frac{1}{\rho} \frac{d}{dz} \left(kmvis \frac{d\vec{v}}{dz} \right). \quad (\text{B2})$$

Since WACCM-X uses pressure coordinates we use the chain rule to rewrite the horizontal gradient term

$$\nabla_p \vec{v} = \nabla_z \vec{v} + \left(\frac{\partial p}{\partial z} \right) (\nabla_p z) \left(\frac{\partial \vec{v}}{\partial p} \right), \quad (\text{B3})$$

and use the hydrostatic relation

$$\frac{\partial p}{\partial z} = -\rho g, \quad (\text{B4})$$

(where g is gravity) to rewrite the horizontal molecular viscosity equation

$$\nabla_z \vec{v} = \nabla_p \vec{v} + \rho \nabla_p \Phi \left(\frac{\partial \vec{v}}{\partial p} \right), \quad (\text{B5})$$

600 where $\Phi = gz$ is the geopotential. The second term on the right-hand side of (B5) is
 601 found to be much smaller than the first throughout the computational domain, and has
 602 been neglected in the current implementation. Hence the operators are applied along pres-
 603 sure surfaces rather than z surfaces. For the vertical a transformation to pressure coord-
 604 inates yields

$$\frac{d\vec{v}}{dt} = g^2 \frac{d}{dp} \left(kmvis \rho \frac{d\vec{v}}{dp} \right). \quad (\text{B6})$$

605 A unified infrastructure has been implemented so that the dynamical core ("horizontal")
 606 and physics package (vertical) fetch the viscosity coefficient (*kmvis*) coefficients from the
 607 same code module.

The frictional dissipation of kinetic energy is ultimately turned into heating at the molecular scale. In WACCM-X the change in kinetic energy due to molecular viscosity is turned into heat at each grid point

$$\Delta T_{heat} = -\frac{1}{c_p} \left(K^{(new)} - K^{(old)} \right) \quad (\text{B7})$$

608 where $K^{(old)} = \frac{1}{2} \vec{v}^2$ is specific kinetic energy before applying the molecular viscosity
 609 operator to \vec{v} and $K^{(new)}$ is after. While this guarantees a closed energy budget it is not
 610 entirely correct since the kinetic energy equation terms associated with molecular vis-
 611 cosity have a dissipative and a diffusive term; only the dissipative term should be added
 612 as heating (Bister & Emanuel, 1998). A way to do this rigorously and consistently in spher-
 613 ical geometry is presented in Becker and Burkhardt (2007).

The thermal conductivity equations take the same form as molecular viscosity except for a c_p term in equations

$$\frac{d}{dt} (c_p T) = \frac{1}{\rho} \nabla_z (kmcnd \nabla_z T), \quad (\text{B8})$$

and

$$\frac{d}{dt} (c_p T) = \frac{1}{\rho} \frac{d}{dz} \left(kmcnd \frac{dT}{dz} \right). \quad (\text{B9})$$

614 where *kmcnd* is the conductivity coefficient that is a function of the major species. The
 615 transformation to pressure coordinates is the same as outlined above and not repeated
 616 here.

617 **Appendix C Vertical profiles for artificial viscosity (sponge layer damp-** 618 **ing)**

619 It was found challenging to stabilize WACCM-X likely due to the less diffusive char-
 620 acteristics of the spectral-element dynamical core (compared to the finite-volume dynam-
 621 ical core) and perhaps excessively large tendencies from the parameterizations (the lat-
 622 ter was not investigated in detail).

623 The spectral-element dynamical core by default uses constant hyperviscosity (∇_{ψ}^4)
 624 applied to all prognostic variables $\psi = T$ (temperature), $\psi = \Delta p$ (pressure-level thick-
 625 ness), divergence $\psi = \delta$ and vorticity $\psi = \zeta$. For a nominal 1° resolution the constant
 626 reference values for fourth-order hyperviscosity are

$$\begin{aligned} \nu_\delta^{(ref)} &= 2.5 \times 10^{15} \text{ m}^4/\text{s}, \\ \nu_\zeta^{(ref)} &= 0.5 \times 10^{15} \text{ m}^4/\text{s}, \\ \nu_T^{(ref)} = \nu_{\Delta p}^{(ref)} &\equiv \nu_{\Delta x=1^\circ}^{(ref)} = 1.0 \times 10^{15} \text{ m}^4/\text{s}, \end{aligned} \quad (\text{C1})$$

(reference hyperviscosity coefficients for nominally 1° resolution; $\Delta x^{(ref)} \equiv 110km$)

627 and are scaled for other resolutions using

$$\begin{aligned}\nu_\delta^{(ref)} &= 2.5 \times \Upsilon, \\ \nu_\zeta^{(ref)} &= 0.5 \times \Upsilon, \\ \nu_T^{(ref)} = \nu_{\Delta p}^{(ref)} &= 1.0 \times \Upsilon,\end{aligned}\tag{C2}$$

(reference hyperviscosity coefficients for any resolution)

where

$$\Upsilon \equiv \nu^{(scaling)} \left[\left(\frac{30}{ne} \right) \nu_{\Delta x=1^\circ}^{(ref)} \right]^\lambda\tag{C3}$$

and

$$\nu^{(scaling)} = \left[\frac{R}{R^{(Earth)}} \right] \nu_{\Delta x=1^\circ}^{(ref)} \left[\Delta x^{(ref)} \right]^{-\lambda}\tag{C4}$$

628 where $\lambda \equiv \frac{1}{\log_{10} 2}$ is the scaling coefficient (which ensures that viscosity coefficients de-
629 crease a logarithmic decade for a doubling of resolution), $\nu_{\Delta x=1^\circ}^{(ref)}$ is a reference value for
630 viscosity at 1° resolution given in (C2) and associated average grid spacing $\Delta x^{(ref)} =$
631 $110km$. Since spectral-elements is also run on other planets the viscosity coefficients need
632 to be scaled accordingly. Hence we have introduced R , which is the mean radius of the
633 planet in question, and $R^{(Earth)}$ is the mean radius of Earth. The resolution is speci-
634 fied in terms of number of elements along a cubed-sphere side, ne . For 1° $ne = 30$ and
635 for 0.25° $ne = 120$ and the resulting viscosity coefficients are $\nu_\delta^{(ref)} = 2.5 \times 10^{13}$, $\nu_\zeta^{(ref)} =$
636 0.5×10^{13} and $\nu_T^{(ref)} = \nu_{\Delta p}^{(ref)} = 1.0 \times 10^{13}$. Some of these values are adjusted for
637 WACCM-X:

$$\begin{aligned}\nu_\delta^{(ref)} &= 1.5 \times 10^{13} m^4/s, \\ \nu_\zeta^{(ref)} &= 1.0 \times 10^{13} m^4/s, \\ \nu_T^{(ref)} = \nu_{\Delta p}^{(ref)} &= 1.0 \times 10^{13} m^4/s,\end{aligned}$$

638 these values were chosen empirically to maintain numerical stability.

In addition there is increased Laplacian damping (∇^2) near the model top using the following coefficient

$$\mu_\psi(k) = \Gamma^{(\mu)}(k) \mu^{(ref)}, \text{ where } \psi = T, \vec{v}, \Delta p,\tag{C5}$$

if $\Gamma^{(\mu)}(k) > 0.15$ else $\mu^{(\mu)}(k) = 0$. The reference value $\mu^{(ref)}$ is specified via the namelist variable `se_nu_top` (for WACCM-X $\mu^{(ref)} = 1E6m^2/s$ and for any other configuration it is $\mu^{(ref)} = 1.25E5m^2/s$). The scaling function $\Gamma^{(\mu)}$ is given by

$$\Gamma^{(\mu)}(k) = 8 \left\{ 1 + \tanh \left[\log \left(\frac{p_{top}}{p_k} \right) \right] \right\},\tag{C6}$$

639 (Lauritzen et al., 2011) where p_{top} is the pressure at the model top and p_k is the mid pres-
640 sure in level k .

Unfortunately this level of damping was found insufficient for stabilizing WACCM-X. Even drastic increases in $\mu^{(ref)}$ were found unsuccessful in terms of stability. By trial and error stability was achieved by increasing ∇^4 damping in the top model layers using the following scaling function

$$\Gamma^{(\nu)}(k) = \frac{1}{2} \left\{ 1 + \tanh \left[2 \log \left(\frac{p_{k_s}}{p_k} \right) \right] \right\}\tag{C7}$$

where k_s is the mid-point of the sponge given by namelist variable

$$k_s = \text{se_sponge_del4_lev}.\tag{C8}$$

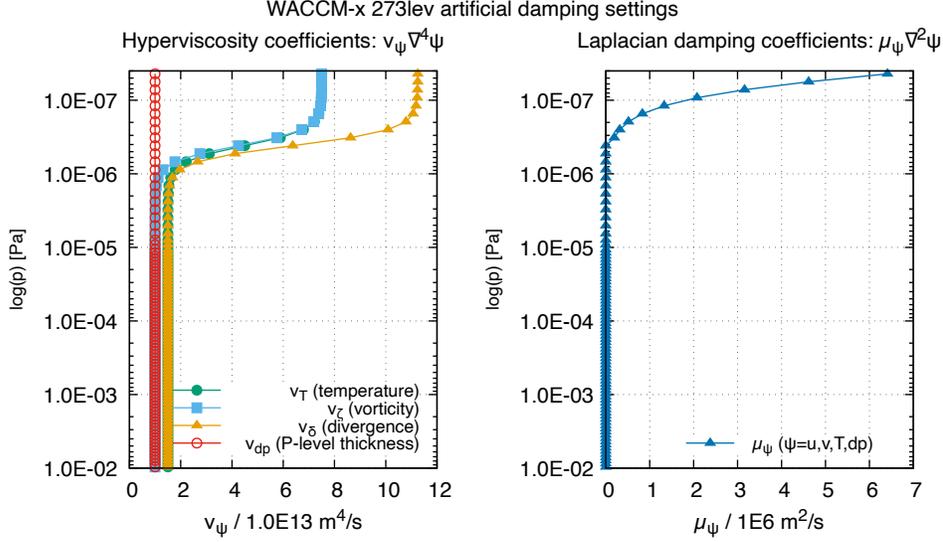


Figure C1: Left plot shows the hyperviscosity coefficient (ν_ψ , $\psi = \delta, \zeta, T, \Delta p$) in units of $1 \times 10^{13} m^4/s$ as a function of pressure (Pa). The right Figure is the Laplacian sponge layer diffusion coefficient (μ_ψ , $\psi = \vec{v}, T, \Delta p$) in units of $1E6 m^2/s$ as a function of pressure. The artificial Laplacian damping coefficient is much smaller than the physical molecular viscosity and thermal conductivity coefficients.

For WACCM-X we use $k_s = 10$ (for model tops up to 140km we use $k_s = 3$). The following damping coefficients, ν_ψ , are used for ∇^4 operator on divergence (δ), vorticity (ζ) and temperature (T)

$$\nu_\psi(k) = [1 - \Gamma^{(\nu)}(k)] \nu_\psi^{(ref)} + \Gamma^{(\nu)}(k) \nu_\psi^{(max)}, \quad \text{where } \psi = \delta, \zeta, T \quad (C9)$$

641 where the maximum damping coefficients (for model tops above ~ 42 km, i.e. WACCM
642 and WACCM-X)¹ are

$$\nu_\delta^{(max)} = 7.5 \nu_{\Delta p}^{(ref)}, \quad \text{namelist se_sponge_del4_nu_div_fac}=7.5, \quad (C10)$$

$$\nu_\zeta^{(max)} = 5.0 \nu_{\Delta p}^{(ref)}, \quad \text{namelist se_sponge_del4_nu_fac}=5.0, \quad (C11)$$

$$\nu_T^{(max)} = 5.0 \nu_{\Delta p}^{(ref)}, \quad \text{namelist se_sponge_del4_nu_fac}=5.0, \quad (C12)$$

643 The damping coefficients as a function of pressure are shown in Figure C1

644 Acknowledgments

645 This work is partially funded by the NCAR System for Integrated Modeling of the At-
646 mosphere (SIMA) project. HLL acknowledges partial support by NASA Grants 80NSSC20K1323,
647 80NSSC20K0601, 80NSSC20K0633, and 80NSSC21K1305. National Center for Atmo-
648 spheric Research is a major facility sponsored by the National Science Foundation un-
649 der Cooperative Agreement No. 1852977. Numerical simulations were performed on NWSC/NCAR
650 Cheyenne Supercomputers with computing resources provided by the NCAR Strategic
651 Capability (NSC) allocation and the Computational and Information Systems Labora-
652 tory (CISL) at NCAR. NCAR CESM/WACCM is an open-source community model, and

¹ for model configurations with model tops below ~ 42 km `se_sponge_del4_nu_div_fac=4.5` and `se_sponge_del4_nu_div_fac=1.0`

653 is available at <https://doi.org/10.5065/D67H1H0V>. Model output used for this study
 654 is available through GLOBUS (shared end point: <https://tinyurl.com/3hnwjz93>).

655 References

- 656 Akmaev, R. A. (2011). Whole atmosphere modeling: Connecting terrestrial and
 657 space weather. *Rev. Geophys.*, *49*. doi: 10.1029/2011RG000364
- 658 Azeem, I., Walterscheid, R. L., & Crowley, G. (2018). Investigation of acoustic waves
 659 in the ionosphere generated by a deep convection system using distributed net-
 660 works of gps receivers and numerical modeling. *Geophysical Research Letters*,
 661 *45*, 8014-8021. doi: 10.1029/2018GL078107
- 662 Azeem, I., Yue, J., Hoffmann, L., Miller, S. D., Straka, W. C., & Crowley, G. (2015).
 663 Multisensor profiling of a concentric gravity wave event propagating from the
 664 troposphere to the ionosphere. *Geophysical Research Letters*, *42*, 7874–7880.
 665 (2015GL065903) doi: 10.1002/2015GL065903
- 666 Becker, E., & Burkhardt, U. (2007). Nonlinear horizontal diffusion for GCMs.
 667 *Monthly Weather Review*, *135*(4), 1439 - 1454. Retrieved from [https://](https://journals.ametsoc.org/view/journals/mwre/135/4/mwr3348.1.xml)
 668 journals.ametsoc.org/view/journals/mwre/135/4/mwr3348.1.xml doi:
 669 10.1175/MWR3348.1
- 670 Becker, E., & Vadas, S. E. (2018). Secondary gravity waves in the winter meso-
 671 sphere: Results from a high-resolution global circulation model. *Jour-*
 672 *nal of Geophysical Research: Atmospheres*, *123*, 2605?2627. doi: 10.1002/
 673 2017JD027460
- 674 Becker, E., & Vadas, S. L. (2020). Explicit global simulation of gravity waves in
 675 the thermosphere. *Journal of Geophysical Research: Space Physics*, *125*,
 676 e2020JA028034. doi: <https://doi.org/10.1029/2020JA028034>
- 677 Becker, E., Vadas, S. L., Bossert, K., Harvey, V. L., Zülicke, C., & Hoffmann, L.
 678 (2022). A high-resolution whole-atmosphere model with resolved gravity waves
 679 and specified large-scale dynamics in the troposphere and stratosphere. *Jour-*
 680 *nal of Geophysical Research: Atmospheres*, *127*(2), e2021JD035018.
- 681 Bister, M., & Emanuel, K. A. (1998). Dissipative heating and hurricane inten-
 682 sity. *Meteorology and Atmospheric Physics*, *65*, 233?240. doi: [https://doi.org/](https://doi.org/10.1007/BF01030791)
 683 [10.1007/BF01030791](https://doi.org/10.1007/BF01030791)
- 684 Borchert, S., Zhou, G., Baldauf, M., Schmidt, H., Zängl, G., & Reinert, D. (2019).
 685 The upper-atmosphere extension of the icon general circulation model (version:
 686 ua-icon-1.0). *Geoscientific Model Development*, *12*(8), 3541–3569.
- 687 Catling, D. C., & Kasting, J. F. (2017). *Atmospheric evolution on inhabited and life-*
 688 *less worlds*. Cambridge University Press.
- 689 Codrescu, M., Fuller-Rowell, T., & Foster, J. (1995). On the importance of e-field
 690 variability for joule heating in the high-latitude thermosphere. *Geophysical Re-*
 691 *search Letters*, *22*(17), 2393–2396.
- 692 Emmert, J. T., Picone, J. M., Lean, J. L., & Knowles, S. H. (2004). Global
 693 change in the thermosphere: Compelling evidence of a secular decrease
 694 in density. *Journal of Geophysical Research: Space Physics*, *109*. doi:
 695 10.1029/2003JA010176
- 696 Emmert, J. T., Picone, J. M., & Meier, R. R. (2008). Thermospheric global average
 697 density trends, 1967?2007, derived from orbits of 5000 near-earth objects. *Geo-*
 698 *physical Research Letters*, *35*. doi: 10.1029/2007GL032809
- 699 Ern, M., Preusse, P., Gille, J. C., Hoppelwhite, C. L., Mlynczak, M. G., III,
 700 J. M. R., & Riese, M. (2011). Implications for atmospheric dynamics de-
 701 rived from global observations of gravity wave momentum flux in strato-
 702 sphere and mesosphere. *Journal of Geophysical Research*, *116*. doi:
 703 10.1029/2011JD015821
- 704 Ern, M., Trinh, Q. T., Preusse, P., Gille, J. C., Mlynczak, M. G., Russell III, J. M.,
 705 & Riese, M. (2018). Gracile: a comprehensive climatology of atmospheric grav-

- ity wave parameters based on satellite limb soundings. *Earth System Science Data*, *10*, 857–892. doi: 10.5194/essd-10-857-2018
- Forbes, J. M., Bruinsma, S. L., Doornbos, E., & Zhang, X. (2016). Gravity wave-induced variability of the middle thermosphere. *Journal of Geophysical Research: Space Physics*, *121*, 6914–6923. doi: 10.1002/2016JA022923
- Gasperini, F., Hagan, M. E., & Zhao, Y. (2017). Evidence of tropospheric 90day oscillations in the thermosphere. *Geophysical Research Letters*, *44*, 10,125–10,133. doi: 10.1002/2017GL075445
- Gasperini, F., Liu, H., & McInerney, J. (2020). Preliminary evidence of madden-Julian oscillation effects on ultrafast tropical waves in the thermosphere. *Journal of Geophysical Research: Space Physics*, *125*, e2019JA027649. doi: 10.1029/2019JA027649
- Geller, M. A., Alexander, M. J., Love, P. T., Bacmeister, J., Ern, M., Hertzog, A., ... Zhou, T. (2013). A comparison between gravity wave momentum fluxes in observations and climate models. *Journal of Climate*, *26*, 6383–6405. doi: 10.1175/JCLI-D-12-00545.1
- Gottelman, A., Mills, M. J., Kinnison, D. E., Garcia, R. R., Smith, A. K., Marsh, D. R., ... Randel, W. J. (2019). The whole atmosphere community climate model version 6 (waccm6). *Journal of Geophysical Research: Atmospheres*, *124*, 12380–12403. doi: https://doi.org/10.1029/2019JD030943
- Herrington, A. R., Lauritzen, P. H., Reed, K. A., Goldhaber, S., & Eaton, B. E. (2019). Exploring a lower-resolution physics grid in cam-se-cslam. *Journal of Advances in Modeling Earth Systems*, *11*(7), 1894–1916.
- Herrington, A. R., Lauritzen, P. H., Taylor, M. A., Goldhaber, S., Eaton, B. E., Bacmeister, J. T., ... Ullrich, P. A. (2019). Physics–dynamics coupling with element-based high-order galerkin methods: Quasi-equal-area physics grid. *Monthly Weather Review*, *147*(1), 69–84.
- Huang, C. S., & Kelley, M. C. (1996a). Nonlinear evolution of equatorial spread-f. 1. on the role of plasma instabilities and spatial resonance associated with gravity wave seeding. *Journal of Geophysical Research*, *101*, 283–292.
- Huang, C. S., & Kelley, M. C. (1996b). Nonlinear evolution of equatorial spread-f. 2. gravity wave seeding of rayleigh-taylor instability. *Journal of Geophysical Research*, *101*, 293–302.
- Huba, J. D., & Liu, H.-L. (2020). Global modeling of equatorial spread f with sami3/waccm-x. *Geophysical Research Letters*, *47*, e2020GL088258. doi: 10.1029/2020GL088258
- Hysell, D. L., Kelley, M. C., Swartz, W. E., & Woodman, R. F. (1990). Seeding and layering of equatorial spread f by gravity waves. *Journal of Geophysical Research: Space Physics*, *95*, 17253–17260. doi: 10.1029/JA095iA10p17253
- Jin, H., Miyoshi, Y., Fujiwara, H., Shinagawa, H., Terada, K., Terada, N., ... Saito, A. (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: Space Physics*, *116*. doi: 10.1029/2010JA015925
- John, S. R., & Kumar, K. K. (2012). Timed/saber observations of global gravity wave climatology and their interannual variability from stratosphere to mesosphere lower thermosphere. *Climate dynamics*, *39*(6), 1489–1505.
- Jones Jr., M., Siskind, D. E., Drob, D. P., McCormack, J. P., Emmert, J. T., Dhadly, M. S., ... Jacobi, C. (2020). Coupling from the middle atmosphere to the exobase: Dynamical disturbance effects on light chemical species. *Journal of Geophysical Research: Space Physics*, *125*, e2020JA028331. doi: 10.1029/2020JA028331
- Kelley, M. C., Larsen, M. F., LaHoz, C., & McClure, J. P. (1981). Gravity wave initiation of equatorial spread f: A case study. *Journal of Geophysical Research: Space Physics*, *86*, 9087–9100. doi: 10.1029/JA086iA11p09087

- 761 Krall, J., Huba, J. D., & Fritts, D. C. (2013). On the seeding of equatorial spread f
762 by gravity waves. *Geophysical Research Letters*, *40*, 661–664. doi: 10.1002/grl
763 .50144
- 764 Lauritzen, P. H., Mirin, A., Truesdale, J., Raeder, K., Anderson, J., Bacmeister, J.,
765 & Neale, R. B. (2011). Implementation of new diffusion/filtering operators in
766 the CAM-FV dynamical core. *The International Journal of High Performance*
767 *Computing Applications*. doi: 10.1177/1094342011410088
- 768 Lauritzen, P. H., Nair, R. D., Herrington, A. R., Callaghan, P., Goldhaber, S.,
769 Dennis, J. M., . . . Tribbia, J. J. (2018). Ncar release of cam-se in cesm2.0:
770 A reformulation of the spectral element dynamical core in dry-mass verti-
771 cal coordinates with comprehensive treatment of condensates and en-
772 ergy. *Journal of Advances in Modeling Earth Systems*, *10*, 1537-1570. doi:
773 https://doi.org/10.1029/2017MS001257
- 774 Lauritzen, P. H., Taylor, M. A., Overfelt, J., Ullrich, P. A., Nair, R. D., Goldhaber,
775 S., & Kelly, R. (2017). Cam-se-cslam: Consistent coupling of a conservative
776 semi-lagrangian finite-volume method with spectral element dynamics. *Monthly*
777 *Weather Review*, *145*(3), 833–855.
- 778 Liu, H., Pedatella, N., & Hocke, K. (2017). Medium-scale gravity wave activity in
779 the bottomside f region in tropical regions. *Geophysical Research Letters*, *44*,
780 7099-7105. doi: 10.1002/2017GL073855
- 781 Liu, H.-L. (2007). On the large wind shear and fast meridional transport above
782 the mesopause. *Geophysical Research Letters*, *34*. (L08815) doi: 10.1029/
783 2006GL028789
- 784 Liu, H.-L. (2016). Variability and predictability of the space environment as related
785 to lower atmosphere forcing. *Space Weather*, *14*, 634–658. (2016SW001450)
786 doi: 10.1002/2016SW001450
- 787 Liu, H.-L. (2019). Quantifying gravity wave forcing using scale invariance. *Nature*
788 *Communications*, *10*, 2605. doi: 10.1038/s41467-019-10527-z
- 789 Liu, H.-L. (2021). Effective vertical diffusion by atmospheric gravity waves. *Geo-*
790 *physical Research Letters*, *48*(1), e2020GL091474.
- 791 Liu, H.-L., Bardeen, C. G., Foster, B. T., Lauritzen, P., Liu, J., Lu, G., . . . Wang,
792 W. (2018). Development and validation of the whole atmosphere commu-
793 nity climate model with thermosphere and ionosphere extension (waccm-x
794 2.0). *Journal of Advances in Modeling Earth Systems*, *10*, 381-402. doi:
795 10.1002/2017MS001232
- 796 Liu, H.-L., McInerney, J. M., Santos, S., Lauritzen, P. H., Taylor, M. A., & Pe-
797 datella, N. M. (2014). Gravity waves simulated by high-resolution Whole
798 Atmosphere Community Climate Model. *Geophysical Research Letters*, *41*,
799 9106–9112. (2014GL062468) doi: 10.1002/2014GL062468
- 800 Liu, H.-L., & Roble, R. G. (2002). A study of a self-generated stratospheric
801 sudden warming and its mesospheric/lower thermospheric impacts using
802 coupled TIME-GCM/CCM3. *Journal of Geophysical Research*, *107*. doi:
803 10.1029/2001JD001533
- 804 McClure, J. P., Singh, S., Bamgboye, D. K., Johnson, F. S., & Kil, H. (1998).
805 Occurrence of equatorial f region irregularities: Evidence for tropospheric seed-
806 ing. *Journal of Geophysical Research: Space Physics*, *103*, 29119-29135. doi:
807 10.1029/98JA02749
- 808 Meier, R. (2021). The thermospheric column o/n2 ratio. *Journal of Geophysical Re-*
809 *search: Space Physics*, *126*(3), e2020JA029059.
- 810 Miyoshi, Y., Jin, H., Fujiwara, H., & Shinagawa, H. (2018). Numerical study of
811 traveling ionospheric disturbances generated by an upward propagating gravity
812 wave. *Journal of Geophysical Research: Space Physics*, *123*, 2141-2155. doi:
813 10.1002/2017JA025110
- 814 Nishioka, M., Tsugawa, T., Kubota, M., & Ishii, M. (2013). Concentric waves
815 and short-period oscillations observed in the ionosphere after the 2013

- 816 Moore EF5 tornado. *Geophysical Research Letters*, *40*, 5581–5586. doi:
817 10.1002/2013GL057963
- 818 Okui, H., Sato, K., & Watanabe, S. (2022). Contribution of gravity waves to
819 universal vertical wavenumber ($\sim m^{-3}$) spectra revealed by a gravity-wave-
820 permitting general circulation model. *Journal of Geophysical Research: Atmo-*
821 *spheres*, e2021JD036222.
- 822 Park, J., Lühr, H., Lee, C., Kim, Y. H., Jee, G., & Kim, J.-H. (2014). A climatology
823 of medium-scale gravity wave activity in the midlatitude/low-latitude daytime
824 upper thermosphere as observed by champ. *Journal of Geophysical Research:*
825 *Space Physics*, *119*, 2187–2196. doi: 10.1002/2013JA019705
- 826 Pedatella, N. M., Anderson, J. L., Chen, C. H., Raeder, K., Liu, J., Liu, H.-L.,
827 & Lin, C. H. (2020). Assimilation of ionosphere observations in the whole
828 atmosphere community climate model with thermosphere-ionosphere ex-
829 tension (waccmx). *Journal of Geophysical Research: Space Physics*, *125*,
830 e2020JA028251. doi: <https://doi.org/10.1029/2020JA028251>
- 831 Pedatella, N. M., Fuller-Rowell, T., Wang, H., Jin, H., Miyoshi, Y., Fujiwara, H.,
832 ... Goncharenko, L. (2014). The neutral dynamics during the 2009 sud-
833 den stratosphere warming simulated by different whole atmosphere mod-
834 els. *Journal of Geophysical Research: Space Physics*, *119*, 1306–1324. doi:
835 10.1002/2013JA019421
- 836 Pedatella, N. M., & Liu, H.-L. (2018). The influence of internal atmospheric variabil-
837 ity on the ionosphere response to a geomagnetic storm. *Geophysical Research*
838 *Letters*, *45*(10), 4578–4585.
- 839 Pedatella, N. M., Raeder, K., Anderson, J. L., & Liu, H.-L. (2014). Ensemble data
840 assimilation in the whole atmosphere community climate model. *Journal of*
841 *Geophysical Research*, *119*, 9793–9809. doi: 10.1002/2014JD021776
- 842 Pitteway, M. L. V., & Hines, C. O. (1963). The viscous damping of atmospheric
843 gravity waves. *Canadian Journal of Physics*, *41*, 1935–1948. doi: 10.1139/p63
844 -194
- 845 Qian, L. Y., Solomon, S. C., & Kane, T. J. (2009). Seasonal variation of thermo-
846 spheric density and composition. *Journal of Geophysical Research*, *114*. doi:
847 10.1029/2008JA013643
- 848 Randall, C. E., Harvey, V. L., Singleton, C. S., Bernath, P. F., Boone, C. D.,
849 & Kozyra, J. U. (2006). Enhanced nox in 2006 linked to strong up-
850 per stratospheric arctic vortex. *Geophysical Research Letters*, *33*. doi:
851 10.1029/2006GL027160
- 852 Randall, C. E., Harvey, V. L., Siskind, D. E., France, J., Bernath, P. F., Boone,
853 C. D., & Walker, K. A. (2009). NO_x descent in the Arctic middle atmosphere
854 in early 2009. *Geophysical Research Letters*, *36*. doi: 10.1029/2009GL039706
- 855 Retterer, J. M., & Roddy, P. (2014). Faith in a seed: on the origins of equatorial
856 plasma bubbles. *Annales Geophysicae*, *32*, 485–498. doi: 10.5194/angeo-32-485
857 -2014
- 858 Richmond, A. D. (1995). Ionospheric electrodynamics using magnetic apex coordi-
859 nates. *J. Geomagn. Geoelectr.*, *47*, 191–212.
- 860 Richter, J. H., Glanville, A. A., Edwards, J., Kauffman, B., Davis, N. A., Jaye, A.,
861 ... Oleson, K. W. (2022). Subseasonal earth system prediction with cesm2.
862 *Weather and Forecasting*, *37*, 797 - 815. doi: 10.1175/WAF-D-21-0163.1
- 863 Roble, R. G., & Dickinson, R. E. (1989). How will changes in carbon-dioxide and
864 methane modify the mean structure of the mesosphere and thermosphere. *Geo-*
865 *physical Research Letters*, *16*, 1441–1444.
- 866 Savitzky, A., & Golay, M. J. (1964). Smoothing and differentiation of data by sim-
867 plified least squares procedures. *Analytical chemistry*, *36*(8), 1627–1639.
- 868 Solomon, S. C., Liu, H.-L., Marsh, D. R., McInerney, J. M., Qian, L., & Vitt, F. M.
869 (2018). Whole atmosphere simulation of anthropogenic climate change. *Geo-*
870 *physical Research Letters*, *45*, 1567–1576. doi: 10.1002/2017GL076950

- 871 Stober, G., Kuchar, A., Pokhotelov, D., Liu, H., Liu, H.-L., Schmidt, H., . . . others
872 (2021). Interhemispheric differences of mesosphere–lower thermosphere winds
873 and tides investigated from three whole-atmosphere models and meteor radar
874 observations. *Atmospheric chemistry and physics*, *21*(18), 13855–13902.
- 875 Strickland, D., Evans, J., & Paxton, L. (1995). Satellite remote sensing of thermo-
876 spheric O_2 and solar EUV: 1. theory. *Journal of Geophysical Research: Space*
877 *Physics*, *100*(A7), 12217–12226.
- 878 Swinbank, R., & Ortland, D. A. (2003). Compilation of wind data for the Upper
879 Atmosphere Research Satellite (UARS) Reference Atmosphere Project. *Journal*
880 *of Geophysical Research*, *108*. doi: doi:10.1029/2002JD003135
- 881 Taylor, M. J., Forbes, J., Fritts, D. C., Eckermann, S. D., Snively, J. B., Liu, H.,
882 . . . others (2017). The atmospheric waves experiment (awe): Quantifying the
883 impact of gravity waves on the edge of space. In *Agu fall meeting abstracts*
884 (Vol. 2017, pp. SA31B–01).
- 885 Theurich, G., DeLuca, C., Campbell, T., Liu, F., Saint, K., Vertenstein, M., . . .
886 Dunlap, R. (2016). The earth system prediction suite: Toward a coordinated
887 u.s. modeling capability. *Bulletin of the American Meteorological Society*, *97*,
888 1229 - 1247. doi: 10.1175/BAMS-D-14-00164.1
- 889 Vadas, S. L. (2007). Horizontal and vertical propagation and dissipation of gravity
890 waves in the thermosphere from lower atmospheric and thermospheric sources.
891 *Journal of Geophysical Research*, *112*. doi: doi:10.1029/2006JA011845
- 892 Vadas, S. L., & Becker, E. (2019). Numerical modeling of the generation of tertiary
893 gravity waves in the mesosphere and thermosphere during strong mountain
894 wave events over the southern andes. *Journal of Geophysical Research: Space*
895 *Physics*, *124*, 7687-7718. doi: 10.1029/2019JA026694
- 896 Vadas, S. L., & Fritts, D. C. (2001). Gravity wave radiation and mean responses to
897 local body forces in the atmosphere. *Journal of the Atmospheric Sciences*, *58*,
898 2249-2279.
- 899 Vadas, S. L., & Liu, H.-L. (2013). Numerical modeling of the large-scale neutral and
900 plasma responses to the body forces created by the dissipation of gravity waves
901 from 6 h of deep convection in Brazil. *Journal of Geophysical Research*, *118*,
902 2593-2617. doi: 10.1002/jgra.50249
- 903 Verronen, P. T., Andersson, M. E., Marsh, D. R., Kovács, T., & Plane, J. M. C.
904 (2016). Waccm-d?whole atmosphere community climate model with d-region
905 ion chemistry. *Journal of Advances in Modeling Earth Systems*, *8*, 954-975.
906 doi: 10.1002/2015MS000592
- 907 Wang, H., Fuller-Rowell, T. J., Akmaev, R. A., Hu, M., Kleist, D. T., & Iredell,
908 M. D. (2011). First simulations with a whole atmosphere data assimilation and
909 forecast system: The January 2009 major sudden stratospheric warming. *Journal*
910 *of Geophysical Research: Space Physics*, *116*. doi: 10.1029/2011JA017081
- 911 Watanabe, S., & Miyahara, S. (2009). Quantification of the gravity wave forcing
912 of the migrating diurnal tide in a gravity wave-resolving general circulation
913 model. *Journal of Geophysical Research: Atmospheres*, *114*. (D07110) doi:
914 10.1029/2008JD011218
- 915 Wu, J., Feng, W., Liu, H.-L., Xue, X., Marsh, D. R., & Plane, J. M. C. (2021). Self-
916 consistent global transport of metallic ions with waccm-x. *Atmospheric Chem-*
917 *istry and Physics*, *21*, 15619–15630. doi: 10.5194/acp-21-15619-2021