RESEARCH ARTICLE

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Key Points:

- Unusually large thermospheric effects for a modest geomagnetic storm
- TADs have a strong latitude, longitude, and altitude dependence
- Finding two important thermospheric storm characteristics

Supporting Information:

- Readme
- Movie S1

Correspondence to:

G. Lu,
ganglu@ucar.edu

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Global ionospheric and thermospheric response to the 5 April 2010 geomagnetic storm: An integrated data-model investigation

G. Lu¹, M. E. Hagan¹, K. Häusler¹, E. Doornbos², S. Bruinsma³, B. J. Anderson⁴, and H. Korth⁴

¹High Altitude Observatory, National Center for Atmospheric Research, Boulder, Colorado, USA, ²Aerospace Engineering, Delft University of Technology, Delft, Netherlands, ³Department of Terrestrial and Planetary Geodesy CNES, Toulouse, France, ⁴The Johns Hopkins University Applied Physics Laboratory, Laurel, Maryland, USA

Abstract We present a case study of the 5 April 2010 geomagnetic storm using observations and numerical simulations. The event was driven by a fast-moving coronal mass ejection and despite being a moderate storm with a minimum *Dst* near -50 nT, the event exhibited elevated thermospheric density and surges of traveling atmospheric disturbances (TADs) more typically seen during major storms. The Thermosphere-Ionosphere-Mesosphere-Electrodynamics General Circulation Model (TIMEGCM) was used to assess how these features were generated and developed during the storm. The model simulations gave rise to TADs that were highly nonuniform with strong latitude and longitude/local time dependence. The TAD phase speeds ranged from 640 m/s to 780 m/s at 400 km and were $\sim 5\%$ lower at 300 km and approximately 10–15% lower at 200 km. In the lower thermosphere around 100 km, the TAD signatures were nearly unrecognizable due to much stronger influence of upward propagating atmospheric tides. The thermosphere simulation results were compared to observations available from the Gravity Field and Steady-State Ocean Circulation Explorer (GOCE), CHALLENGING Minisatellite Payload (CHAMP) and Gravity Recovery and Climate Experiment (GRACE) satellites. Comparison with GOCE data shows that the TIMEGCM reproduced the cross-track winds over the polar region very well. The model-data comparison also revealed some differences, specifically, the simulations underestimated neutral mass density in the upper thermosphere above ~ 300 km and overestimated the storm recovery time by 6 h. These discrepancies indicate that some heating or circulation dynamics and potentially cooling processes are not fully represented in the simulations, and also that updates to some parameterization schemes in the TIMEGCM are warranted.

1. Introduction

A coronal mass ejection (CME) was unleashed from the Sun on 3 April 2010 and arrived at the Earth 2 days later [Möstl *et al.*, 2010]. The fast-moving CME drove an interplanetary shock ahead of the ejecta. The interplanetary shock interacted with the Earth's magnetosphere and triggered the onset of a geomagnetic storm at 08:27 UT on 5 April 2010, which was followed by a series of magnetic intensifications over the period of 5–7 April as the CME ejecta or interplanetary coronal mass ejection (ICME) passed across the Earth. It was the first notable geomagnetic storm of solar cycle 24. Despite being a relatively weak storm as gauged by the *Dst* index (the minimum *Dst* was about -50 nT on 5 April), it nevertheless had some devastating space weather impacts, including the malfunction of the Galaxy 15 communication satellite [Allen, 2010] and widespread GPS scintillations ranging from the Arctic to Antarctic [Prikryl *et al.*, 2011; Kinrade *et al.*, 2012; Sieradzki *et al.*, 2013]. Galaxy 15 was in geostationary orbit at an altitude of 35,785 km. The satellite anomaly started at 09:48 UT on 5 April, about 80 min after the storm commencement, and the spacecraft did not regain normal operation until 3 years later in April 2013. While the exact nature of the satellite anomaly is still unclear, it has been suggested that the significant increase in energetic particle fluxes in the premidnight sector at geosynchronous orbit might have caused surface and internal charging of the satellite [Allen, 2010; Denig *et al.*, 2010; Connors *et al.*, 2011; Clilverd *et al.*, 2012]. The detrimental space weather effects on telecommunication and space assets beg for further understanding of the physical processes responsible for producing storm-time disturbances in order to improve our ability to specify, and eventually predict, the state of the near-Earth space environment.

During geomagnetic storms, strong electric fields and currents are transmitted between the magnetosphere and the high-latitude ionosphere, producing enhanced Joule heating and auroral particle precipitation in the

auroral zone. The conductivity of the ionosphere increases, neutral winds are accelerated, the thermosphere is heated and its composition is modified, and ionospheric plasma convection is intensified and highly distorted. The perturbed neutral winds and composition propagate equatorward, creating ionospheric and thermospheric disturbances over the entire globe [e.g., *Prölss*, 1993; *Fuller-Rowell et al.*, 1994, 1997; *Rishbeth*, 1989]. Additional variability of the ionosphere and thermosphere is produced by variations in the tides and planetary waves that propagate upward from below. The dynamics, electrodynamics, energetics, and chemistry of the coupled magnetosphere-ionosphere-thermosphere (MIT) system are strongly interactive such that complex physics-based models are needed to understand the underlying dynamics and drivers of these disturbances.

Despite extensive efforts in observations, theory, and numerical modeling, comprehensive understanding and accurate descriptions of thermospheric and ionospheric response to geomagnetic storms are far from complete. This investigation aims to shed new light on the complex behavior of the MIT system during storm conditions by carrying out detailed numerical simulations using the Thermosphere-Ionosphere-Mesosphere-Electrodynamics General Circulation Model (TIMEGCM) developed at the National Center for Atmospheric Research (NCAR) subject to a comprehensive specification of high-latitude electrodynamic forcing derived from the latest, global, data sets available. The study is also motivated by the unprecedented thermospheric observations available from three satellites during the event: Gravity Field and Steady-State Ocean Circulation Explorer (GOCE) at an average altitude of 270 km in the near dawn-dusk meridional plane, CHALLENGING Minisatellite Payload (CHAMP) at about 300 km in the noon-midnight meridional plane, and Gravity Recovery and Climate Experiment (GRACE) satellites at about 474 km in the 8:30–20:30 meridional plane. The suite of thermospheric observations provides direct observational tests of the simulated thermospheric response. An overarching goal of this study is to validate the first-principle TIMEGCM model and to identify improvements that may be needed to provide better predictions of space weather effects in the ionosphere and thermosphere.

2. Data and Model Description

CHAMP was a German satellite launched on 15 July 2000 which reentered the Earth's atmosphere on 19 September 2010. CHAMP had a near-circular orbit with an 87.3° inclination. Over the years, its orbital altitude gradually decreased from ~460 km initially to ~300 km toward the end of its mission. During its 10 year mission lifetime, CHAMP generated the first simultaneous high-precision measurements of gravity and magnetic fields. In addition, the highly sensitive accelerometer on board CHAMP, designed to aid in the mapping of the gravity field, also afforded estimates of neutral density and cross-track neutral winds. Compared to density estimates, which have uncertainties of less than 10% under quiet conditions and of 10–40% during geomagnetic storms [*Bruinsma et al.*, 2006], the estimates of cross-track winds are much more sensitive to small errors in the instrument data and models used in the data processing [e.g., *Liu et al.*, 2006; *Sutton et al.*, 2005; *Doornbos et al.*, 2010]. During April 2010, CHAMP cross-track wind data are considered inaccurate, most likely because of the loosened thermal control at the accelerometer, resulting from degradation of the batteries near the end of the satellite lifetime. For this reason, CHAMP wind data have not been used during this study.

GRACE is a joint U.S./German mission launched on 17 March 2002, consisting of two identical spacecraft flying about 220 km apart in an 89° inclination orbit. The spacecraft have been providing measurements for the determination of temporal variations in the Earth's gravity field and thermospheric mass density at an altitude above 470 km. The GRACE accelerometers are an order of magnitude more precise than the CHAMP accelerometer, allowing accurate determination of the much lower in-track neutral density at this higher altitude. However, because of this higher orbital altitude, the increased relative influence of solar and Earth radiation pressure accelerations caused by the reduced air density has prevented the derivation of accurate winds from the GRACE cross-track accelerometer measurements.

GOCE was launched on 17 March 2009 and reentered into the lower atmosphere on 11 November 2013. The satellite was dedicated to measuring the Earth's gravitational field in great detail. GOCE was in a 96.7° inclination, sunsynchronous near-circular dawn-dusk orbit at ~270 km. Among the instruments on board GOCE were three pairs of three-axis accelerometers. These accelerometers were similar in design to those on

Figure 1. Distributions of (a) ACE solar wind bulk speed, (b) the IMF B_y and B_z components in GSM coordinates, (c) the 1 min Dst index derived from 53 midlatitude and low-latitude stations in black, and the standard Dst in red, (d) the 1 min AE index derived from 88 auroral stations, and (e) the 5 min polar cap potential drop in the Northern Hemisphere on 5 April 2010. A time shift of 32 min has been applied to the ACE data to account for the solar wind propagation from its upstream location to the Earth. The time resolution is 64 s for the ACE solar wind data and 16 s for the ACE magnetometer data.

board the CHAMP and GRACE spacecraft but with much improved stability and accuracy [Floborghagen *et al.*, 2011]. The GOCE accelerometers provided high-quality measurements of in-track neutral mass density [Doornbos *et al.*, 2014; Bruinsma *et al.*, 2014] and cross-track neutral winds.

The Active Magnetosphere and Planetary Electrodynamics Response Experiment (AMPERE) experiment [Anderson *et al.*, 2014] uses magnetic field data from the Iridium Communications satellite constellation. The constellation consists of more than 66 polar-orbiting satellites in circular polar orbits with 780 km altitude and 86° inclination, configured in six orbit planes equally spaced in longitude. There are 11 satellites in each orbit plane, not including on-orbit spares, distributed evenly along each orbit. Data are processed in the AMPERE Science Data Center to derive perturbations dominated by Birkeland current signals. In this study, the 19.4 s resolution AMPERE data, corresponding

to a latitudinal resolution of about 1.3°, were used as one of the inputs to the assimilative mapping of ionospheric electrodynamics (AMIE) procedure.

The AMIE data assimilation algorithm was first developed at NCAR in 1988 [Richmond and Kamide, 1988] and has since undergone continuous improvement in spatial resolution and its ability to ingest new types of data [e.g., Lu *et al.*, 1998, 2001a]. The objective of the AMIE procedure is to obtain optimal estimates of high-latitude ionospheric electrodynamic fields by combining various direct and indirect observations of these fields. Along with the ionospheric convection patterns, distributions of height-integrated ionospheric horizontal currents, field-aligned currents at the top of the ionosphere, auroral energy flux and characteristic energy, and Joule heating are also by-products of AMIE. For this particular event, the data input to AMIE includes magnetic field measurements from AMPERE, ion drifts from the DMSP satellites (e.g., F15-18), and the SuperDARN radar network in both hemispheres, precipitating auroral electron fluxes from DMSP and ground magnetic perturbations from 203 magnetometer stations worldwide.

The TIMEGCM [Roble and Ridley, 1994; Roble, 1995] is a global model of the mesosphere-thermosphere-ionosphere system, extending from ~30 km to 500–800 km altitude (depending on solar activity). The model incorporates aeronomical, dynamical, and electrodynamic processes appropriate for these regions. The lower boundary of the TIMEGCM is specified by climatological tides based on the Global Scale Wave Model (GSWM-02) [Hagan and Forbes, 2002, 2003]. In addition, the 24 h averaged values of temperature, zonal and meridional winds, and geopotential height at 10 hPa (~30 km) from the European Centre for Medium-range Weather Forecasts reanalysis are used to represent other longer-period dynamical forcing generated below the TIMEGCM lower boundary. The upper boundary inputs to the model include the solar UV and EUV fluxes as parameterized based on the $F_{10.7}$ index together with auroral particle precipitation and plasma drifts derived from AMIE. For this study we used the high-resolution version of the TIMEGCM, which has a horizontal resolution of $2.5^\circ \times 2.5^\circ$ in latitude and longitude and a vertical resolution of four grid points per scale height with a total of 97 pressure levels. The high-resolution simulation adequately resolves the upward propagating waves and tides as well as the horizontally propagating ionospheric/thermospheric disturbances. The model time step is 1 min, and the model outputs are saved at a 10 min cadence.

Figure 2. (top) The IMF B_y and B_z components in GSM coordinates during the period from 04 to 16 UT on 5 April 2010, and the tick marks are 30 min apart. (a–f) The subsequent dial plots show the AMPERE magnetic field perturbations over the northern polar region at selected UT times as marked in Figure 2 (top). Different color vectors represent measurements from the different Iridium satellites. The center of the dial plot corresponds to the magnetic North Pole.

3. Results

3.1. Geophysical Conditions

The geophysical conditions on 5 April 2010 are displayed in Figure 1. Figures 1a and 1b show the solar wind bulk speed and the interplanetary magnetic field (IMF), respectively. The Dst index (solid line in Figure 1c) is derived from 53 stations located below 40° magnetic latitude (MLAT) with temporal resolution of 1 min, showing a clear signature of a storm sudden commencement (SSC) at 8:27 UT. For comparison, the standard hourly Dst index (red dashed line) obtained from four low-latitude stations does not capture the SSC. The SSC onset was also corroborated by observations from the Interstellar Boundary Explorer satellite, which showed a sharp increase in energetic neutral atom emissions from the dayside magnetopause at $08:26:17 \pm 9$ s UT [McComas *et al.*, 2012]. The AE index in Figure 1d was derived from 88 ground magnetometer stations located between 55° and 77° MLAT in the Northern and Southern Hemispheres. Figure 1e plots the cross-polar cap potential drop in the Northern Hemisphere as derived from the AMIE procedure. The solar wind and IMF data from the ACE spacecraft have been time-shifted by 32 min to account for solar wind propagation from the upstream location to the dayside magnetopause. This time shift is estimated based on the time difference between the solar wind pressure pulse associated with the jump in solar wind speed and the onset of SSC. The method to determine the time delay has been described in more detail in Lu *et al.* [2011].

The storm was associated with a fast ICME traveling at a speed exceeding 700 km/s, which produced a compressed sheath region (as marked by the vertical dashed lines) in front of the ICME. Within the sheath region, the IMF was highly fluctuating, flapping from southward to northward. But the magnetic field inside the ICME was primarily westward. Consequently, the storm was initially triggered by the proceeding sheath field rather than the ICME itself. Although it was a weak storm according to the Dst index [Gonzalez *et al.*, 1994], which had a minimum value of around -50 nT, the magnitudes of both AE (with a maximum value of ~ 3000 nT) and the cross-polar cap potential drop (a maximum value of 240 kV) are comparable to those

Figure 3. Consecutive ionospheric convection patterns from 07:40 to 14:00 UT in the Northern Hemisphere plotted in magnetic latitude versus magnetic local time, with a contour interval of 10 kV. Blue contours represent clockwise plasma convection, and red contours counter clockwise convection.

typically seen during major storms such as the Bastille Day storm on 15–16 July 2000 and the Halloween storm on 30–31 October 2003 [Lu, 2006]. This is an interesting feature of the storm, and we speculate that it may be attributed to the fact that the ICME magnetic field was predominantly orientated in the east-west direction instead of southward, such that the auroral electrodynamics were strongly driven but without the intense ring current injections associated with major storms. Here we focus on the ionospheric and thermospheric responses to the storm and leave examination of the magnetospheric particle injection physics for this storm to future work.

3.2. Ionospheric Response

Figure 2 depicts the AMPERE magnetic field perturbations before and during the passage of the ICME. At 08:00 UT (Figure 2a), prior to the arrival of the interplanetary shock, the IMF B_z component was weakly northward and B_y was small, the corresponding AMPERE magnetic field perturbations were smaller than 100 nT. During the first stage of the storm when IMF B_z was southward and within the sheath, the AMPERE magnetic perturbations were greatly increased, with a maximum magnitude exceeding 800 nT (e.g., Figures 2b and 2c). Figures 2d and 2e correspond to northward B_z but also largely westward B_y ($B_y \approx -10$ nT) still within the sheath region and illustrate large AMPERE magnetic field perturbations over the polar region, particularly over the dayside cusp near local noon around 75° – 80° MLAT. Inside the ICME at 1400 UT when the IMF was predominantly westward, strong westward magnetic perturbations were present from postnoon to dawn (see Figure 2f), indicating highly skewed field-aligned current distributions from the normal dawn-dusk symmetry. A recent paper by Wilder *et al.* [2012] discussed specifically the IMF B_y effects on Joule heating intensification over the polar cap during the same event, which subsequently caused a thermospheric density enhancement.

Figure 3 shows the global ionospheric convection patterns derived from AMIE based on the AMPERE magnetic field data together with measurements from SuperDARN, DMSP, and ground magnetometers. The patterns are shown at a 20 min cadence from 07:40 UT to 14:00 UT. Over this 6 h interval, the large-scale

plasma convection in the ionosphere evolved from a weak two-cell pattern to a strong two-cell pattern, to a highly distorted pattern with multiple cells, and then to a skewed two-cell pattern. The figure clearly demonstrates the very dynamic reconfiguration of global ionospheric convection in response to the rapid IMF changes associated with the ICME. Understanding how thermospheric neutral density and winds respond to such dynamic magnetospheric forcing is the primary focus of this study, which is discussed in the following sections.

3.3. Thermospheric Response

Figure 4 shows maps of $\mathbf{E} \times \mathbf{B}$ drifts and neutral winds from the TIMEGCM at selected UT times when the GOCE satellite passed over the northern polar cap. These maps are plotted in geographic coordinates poleward of 50°N , and the center of the map is the geographic North Pole. The plots show the neutral winds at 270 km altitude, corresponding to the average GOCE altitude. Note that, for clarity, wind vectors are plotted on a much coarser grid than the model's actual computational grid (e.g., 2.5° in latitude and longitude). The red arrows are the cross-track winds measured by GOCE. During the six consecutive GOCE passes, the ion drift undergoes dramatic changes in response to the passage of the turbulent sheath region and the subsequent CME ejecta. Although the neutral winds also change with time, they do not follow the ion motion because of the neutral atmosphere's large inertia as well as forcings other than ion drag. At 07:51 UT, prior to the arrival of the ICME, the neutral winds are similar to the ion drifts over most of the

Figure 4. Maps of (left column) $\mathbf{E} \times \mathbf{B}$ drifts and (right column) neutral winds from the TIMEGCM at 270 km in the Northern Hemisphere plotted in geographic latitude versus local time. The marked UT in each panel corresponds to the time when the GOCE spacecraft was closest to the North Pole. The GOCE cross-track winds are plotted in red arrows.

Figure 5. (top row) GOCE neutral mass density measured at (left column) dawn and (right column) dusk. (middle row) Modeled neutral density extracted along the satellite track. (bottom row) Modeled neutral density at fixed local times and a fixed altitude of 266 km. The vertical axis is in geographic latitude. The dashed lines in Figure 5 (bottom row) highlight the main TADs on 5 April with the arrows indicating the TAD propagation directions. The apparent hemispheric difference in Figure 5 (top and middle rows) is due to GOCE's slightly elliptical orbit, which had an average altitude of 266 km in the Northern Hemisphere and 274 km in the Southern Hemisphere.

Northern Hemisphere. But in the polar region above 80°N, the winds exceed the $\mathbf{E} \times \mathbf{B}$ drifts owing to the rather weak magnetospheric forcing under the prevailing weakly northward IMF condition. There is a factor of 2 increase in the GOCE winds in the polar cap from 07:51 UT to 09:20 UT after IMF B_z turned from weakly northward to southward. From 09:20 UT to 13:49 UT when IMF B_y and B_z underwent rapid changes due to the passage of the solar wind sheath region and the CME ejecta, the $\mathbf{E} \times \mathbf{B}$ drifts exhibit very dynamic changes.

Compared to the ion $\mathbf{E} \times \mathbf{B}$ drifts, the response of the neutral winds is relatively subtle in the polar region. The winds along the dawn-dusk meridional plane rotate gradually from antisunward to slightly duskward over the polar region. More dramatic change takes place near dawn below about 65°N where the winds alter from nearly null to sunward and then to antisunward. This simulated neutral wind response is corroborated by the GOCE observations of the cross-track winds from 09:20 to 13:49 UT. During these active times the $\mathbf{E} \times \mathbf{B}$ drift reaches ~1 km/s and the wind speed exceeds 500 m/s in the polar region. A much clearer imitation of the winds to the $\mathbf{E} \times \mathbf{B}$ drifts can be found at 22:48 UT when the magnetic field within the ICME slowly rotated toward a steady value of $B_z \approx B_y \approx -5$ nT. The ion drifts and winds both depict a similar two-cell pattern with the foci located near 03 and 16 LT, respectively. The shift of the flow foci from the dawn-dusk meridional plane is a consequence of the IMF B_y effect.

Thus, despite the rather dynamic IMF conditions and consequently rapid variations in the $\mathbf{E} \times \mathbf{B}$ drifts, we find a good qualitative agreement between GOCE observations and TIMEGCM simulated cross-track winds. In particular, the model does a good job in reproducing the reversal or reduction of the cross-track winds near dusk and dawn. The good quantitative agreement with the GOCE data indicates that realistic $\mathbf{E} \times \mathbf{B}$ drifts such as obtained from AMIE are critically important in specifying thermospheric neutral wind dynamics.

Impulsive storm-time energy dissipation from the magnetosphere perturbs local neutral mass density and temperature in the auroral zone. The perturbed thermospheric structures propagate equatorward under the influence of pressure gradients, forming so-called traveling atmospheric disturbances (TADs). Figure 5 presents the UT-latitude distributions of neutral mass density near the dawn and dusk meridional planes. The top row shows the GOCE measurements of neutral density at ~270 km, and the middle row shows the

Figure 6. Similar to Figure 5 but for comparison with the CHAMP measurements near local noon and local midnight at an average altitude of 301 km. The hemispheric density difference shown in the top and middle rows are due to CHAMP's elliptical orbit.

modeled density sampled along the satellite track over the 3 day period of April 5–7. The color scales are the same for both GOCE data and TIMEGCM simulation results, ranging from 0 to $40 \times 10^{-12} \text{ kg/m}^3$. There is a good agreement between GOCE and TIMEGCM though quantitative differences exist. While storm-induced density changes are obvious, characteristic storm features like TADs are hardly discernible in the neutral density data from GOCE, which was sampling the full-latitude range with an orbital period of 90 min. The bottom row shows the same UT-latitude density distributions from the TIMEGCM using the 10 min cadence model output at all latitudes, instead of the model sampling along the satellite orbit. By using this higher temporal resolution, one can clearly see multiple TADs originating in the polar regions of both hemispheres and propagating toward low latitudes and even into the opposite hemispheres. To help guide reader's eyes, the main TADs on 5 April are highlighted by the dashed lines in Figure 5 (bottom row), with their propagation directions indicated by the arrows. The most prominent TAD shown in Figure 5 (bottom left) originates around 60°N , passes across the equator, and reaches 60°S where it encounters a TAD propagating northward. The time that it takes for the TAD to propagate from 60°N to 60°S is about 5.5 h, indicating a phase speed of $\sim 710 \text{ m/s}$. It is worth noting that the hemispheric difference (e.g., larger mass density in the Northern Hemisphere than in the Southern Hemisphere) shown in Figure 5 (top and middle rows) is due to the fact that GOCE had a slightly elliptical orbit, and its mean altitude was 266 km in the Northern Hemisphere compared to 274 km in the Southern Hemisphere. In Figure 5 (bottom row), however, the mass density from the model is plotted at a fix altitude of 266 km in both hemispheres so that no hemispheric difference is apparent in the plots.

Figure 6 shows the similar data-model comparison with the CHAMP satellite. CHAMP was in a noon-midnight orbital plane at about 300 km. Again, there is a good qualitative agreement between data and model in terms of general morphology. But in this case, the color scale for the modeled neutral density in Figure 6 (middle and bottom rows) is about 20% smaller than that for the CHAMP satellite data in the top panels. Similar to the GOCE measurements shown in Figure 5, TADs are not fully captured by CHAMP, especially on the nightside. However, the TAD features are clearly depicted in Figure 6 (bottom row), making use of the 10 min cadence output of the model at all latitudes simultaneously, at local noon and local midnight. Finally, the data-model comparison of neutral density with the GRACE satellite is shown in Figure 7. GRACE was in a 20–08 LT plane and at an altitude of $\sim 474 \text{ km}$. Note that the color scale for the model output in the bottom two rows is one half of that for the GRACE data in the top row. The TIMEGCM neutral density underestimates illustrated here and in

Figure 7. Similar to Figure 5 but for comparison with the GRACE measurements in the prenoon and premidnight sectors at an average altitude of 474 km. Again, the hemispheric density difference shown in the top and middle rows are due to GRACE's slightly elliptical orbit.

Figure 6 may be attributed to an eddy diffusion coefficient excess [Qian *et al.*, 2009], which is discussed in more detailed in section 4. For this current study, however, we focus on the relative changes associated with the storm so that the shortcomings in the absolute background mass density do not greatly impact our investigation. Indeed, the TIMEGCM is able to reasonably replicate the relative density variations associated with the storm observed by all three satellites.

3.4. Local Time and Altitude Dependence

TADs generated by rapidly varying magnetospheric energy input at high latitudes propagate toward low latitudes and sometimes even into the opposite hemisphere. Because of the nonuniform spatial distributions of magnetospheric energy inputs, thermospheric perturbations during geomagnetic storms exhibit complex latitude and longitude/local time dependence. Figure 8 shows a series of global maps of the height-integrated Joule heating and neutral temperature at 300 km, together with the corresponding horizontal neutral winds, over the course of the storm. These maps are plotted as though they were viewed from a stationary satellite located at local noon and at 60°N. An intensification of Joule heating starts around 08:40 UT over western Greenland, but no noticeable neutral temperature increase can be seen until 09:10 UT. By 10:00 UT, a significant temperature increase is found over the polar region and also in the subauroral zone, away from the most intense Joule heating region. At 11:00 UT, the subauroral band of increased temperature has moved equatorward and westward. At the same time neutral temperature over the polar region has decreased somewhat compared to that at 10:00 UT. At 12:30 UT, Joule heating is reduced to a modest level but the band of high temperature has reached the Southern Hemisphere while corotating with the Earth. A second but rather modest increase in Joule heating takes place between 13:00 and 17:00 UT. Ripples of neutral temperature perturbations are clearly visible at 15:00 UT. Thermospheric temperature continues to be elevated even after the storm has subsided at 23:50 UT.

Though changes in the neutral winds may be less drastic as compared to neutral temperature, the storm does alter the winds globally as also illustrated in Figure 8. At 08:00 UT, prior to the Joule heating enhancement, the winds are weakly westward at midlatitudes and low latitudes, and the largest winds are mostly confined over the polar cap, blowing from dayside to nightside. After the storm onset, the winds undergo large changes both in magnitude and in orientation. In the midlatitude region, the winds are southwestward from 10:00 to 11:00 UT, and turn primarily northwestward with reduced speed at 12:30 UT. After the second but modest

Figure 8. Maps of the height-integrated Joule heating (small global) and neutral temperature at 300 km (big global) based on the coupled AMIE-TIMEGCM simulations on 5 April 2010. Overlain on the temperature maps are the corresponding horizontal wind vectors. The center of each map is set to 60°N at local noon.

Joule heating intensification at 14:10 UT, the midlatitude winds on the dayside vary from southwestward at 15:00 UT to northwestward at 21:00 UT, and then to mostly westward at 23:50 UT. At the same time, winds over the northern polar cap, while remaining mostly toward nightside, swing back and forth between dawnward and duskward. A movie of thermospheric response to Joule heating is included as supporting information, in which the magenta dots indicate the satellite trajectories of GOCE, CHAMP, and GRACE in a 10 min cadence.

To further examine the storm-time neutral wind variability and its longitude/local time and altitude dependence, Figures 9–12 present the UT-latitude distributions of meridional winds at selected LT sectors and altitudes. Similar to the neutral density variations shown in Figures 5–7, TAD features are clearly present in the meridional winds. At 400 km (Figure 9), TADs at local noon and local midnight propagate predominantly from the northern auroral zone into the Southern Hemisphere. But near dawn and dusk, prominent TADs are generated in both hemispheres, and they propagate toward the local magnetic equator where they cross each other and travel further into the opposite hemisphere. The primary TADs in each local time sector are highlighted by the slanted dashed lines in magenta color, along with their respective phase speeds. As shown in Figure 9, TAD's phase speed varies with local time and also differs between the two hemispheres. This is because TADs are a type of gravity wave generated by Joule heating energy dissipation [e.g., Richmond and Matsushita, 1975]. Thus, the phase speed of a TAD is proportional to local sound speed, which in turn depends on local neutral density and temperature. At the altitude of 400 km, the phase speed varies from 640 m/s to 790 m/s, which is comparable to what was found in previous studies based on satellite observations [e.g., Forbes et al., 2005; Bruinsma and Forbes, 2009] and from numerical simulations [e.g., Lu et al., 2001b, 2008]. Bruinsma and Forbes [2009] noted a 10% higher phase speed on the nightside than on the dayside by averaging over 21 storms. Our simulation results, on the other hand, seem to have a slightly higher phase speed on the dayside (787 m/s) as compared with the nightside (762 m/s). This difference is small enough to fall within the error of our estimation and is thus not significant. Bruinsma and Forbes [2009] also concluded that the uncertainty in their estimate is too large to affirm the 10% day-night phase speed difference with certitude. Another noteworthy point is that the phase speed in our simulation is about 10% higher for TADs launched in the Northern Hemisphere than those launched in the Southern Hemisphere. In addition to local sound speed, the background winds also affect the TAD phase speed. As shown by

Figure 9. Simulated meridional winds at 400 km at (top left) local noon, (top right) dawn, (bottom left) midnight, and (bottom right) dusk. The vertical axis is in geographic latitude. The dashed contours represent southward wind, and the solid contours represent northward wind. The thick dashed lines in magenta highlight the phase propagation of the meridional winds.

Balthazor and Moffett [1999], the TAD phase speed increases (decreases) if TADs propagate in the same (opposite) direction as the background winds. Since this event took place post spring equinox, the Northern Hemisphere was more summer-like than the Southern Hemisphere, resulting in a weak southward component in the background neutral winds that counteracts with equatorward propagating TADs in the Southern Hemisphere. The small hemispheric difference in TAD phase speed may be attributed to this seasonal effect. However, such a hemispheric difference was not detected by *Bruinsma and Forbes* [2009].

The TAD features presented in meridional winds at 300 km (Figure 10) are very similar to those at 400 km; however, the corresponding TAD phase speeds are roughly 5% smaller in all local time sectors. Further down in altitude at 200 km (Figure 11), TADs are still easily discernible but their phase speeds become about 10–15% smaller than those at 400 km. At 100 km (Figure 12), there are very little signs of TADs; instead, the meridional winds are strongly modulated by the waves from the lower atmosphere. In the constant local time frame of reference, zonal mean meridional winds and migrating (i.e., sunsynchronous) tidal perturbations are UT/longitude invariant. Thus, the perturbations illustrated in Figure 12 are largely attributable to nonmigrating tides or planetary waves of tropospheric origin. The distorted wave 4 pattern seen at low latitudes at 18 and 24 LT can be primarily attributed to the eastward propagating zonal wave number 3 diurnal (DE3) tide, the eastward propagating zonal wave number 2 semidiurnal (SE2) tide, a stationary planetary wave with zonal wave number 4, or some combination of these three waves. GSWM-02 April meridional tidal climatologies are characterized by the presence of four prominent nonmigrating diurnal components near 100 km, including the DE3 [*Hagan and Forbes*, 2002]. The 100 km GSWM-02 nonmigrating meridional semidiurnal tidal components, including SE2, are individually smaller but their aggregate effects rival those of the prominent GSWM-02 diurnal tides at tropical latitudes [*Hagan and Forbes*, 2003] and suggest that a complicated combination of nonmigrating tides is also present in the 100 km TIMEGCM results. This would explain the distortion of the wave 4 pattern at 18 and 24 LT and the more complicated wave patterns at 6 and 12 LT in Figure 12. A detailed report on the origin, impacts, and variability of thermospheric waves and tides in TIMEGCM before, during, and after the 5 April 2010 solar geomagnetic disturbance is the subject of a follow-on report that is in development.

Figures 9–12 clearly illustrate the fact that magnetospheric forcing attenuates with decreasing altitude. Around 100 km, the influence of storm-time forcing from above is diminished such that the lower thermosphere is largely controlled by forcing from below owing to meteorological weather.

Figure 10. Similar to Figure 8 but for meridional winds at 300 km.

3.5. Global Thermospheric Response

The globally averaged thermospheric response is presented in Figure 13 along with some proxies of prevailing solar geomagnetic conditions. Figure 13a plots the IMF B_y and B_z over the period of 4–8 April. Figure 13b shows the globally integrated Joule heating rate derived from AMIE at a 5 min cadence in the red dotted line, while the solid red line indicates the 3 h running average value. It is interesting to point out the smoothed Joule heating curve closely resembles the reversed hourly Dst index shown in black line, with a correlation coefficient of 0.77 and a time lag of 1 h. However, this resemblance does not imply a causal relationship between Joule heating and Dst since the physical processes that cause Joule heating dissipation into the high-latitude ionosphere and thermosphere are very different from those that bring in particles from the magnetotail and energize them to form enhanced ring current in the inner magnetosphere. Note in this regard that the largest spike in Joule heating at storm onset is not correspondingly reflected in Dst . Figures 13c and 13d show the percent changes of globally averaged neutral temperature and mass density from the TIMEGCM at selected heights. The percent change is referenced to the corresponding mean value on 4 April prior to the storm. There is a clear altitude dependence in the global thermospheric response, with more prominent storm-induced density and temperature enhancement at higher altitudes. This is due to the fact that, although Joule heating peaks in the E region where Pedersen conductivity peaks, Joule heating per unit mass actually increases with altitude due to the exponentially decreasing neutral mass density with altitude. As a result, heating becomes more effective in the upper thermosphere. At 100 km altitude, storm effects become negligible. Figure 13e shows the percent changes of the orbit-averaged measured mass density in solid lines and those from the simulation in dashed lines. Compared to observations, the model underestimates neutral mass density, particularly at the GRACE altitude where the simulated orbitally averaged mass density is only one half of the GRACE observations. This model-data discrepancy was discussed above and also shown in Figure 7.

Figure 13 also reveals two important storm characteristics. The first one concerns thermospheric response time to the storm. As marked by the first vertical dashed line in Figure 13, the initial thermospheric response is nearly simultaneous at all altitudes above 150 km soon after the rapid rise of Joule heating. But the peak of neutral temperature and mass density enhancement lags behind the peak of Joule heating by 3–5 h as indicated by the slanted arrows. Another important characteristic of thermospheric response is the post-storm recovery time. According to *Lei et al.* [2011], a convenient yet useful measure of the recovery time can be defined by the e -folding time after Joule heating energy dissipation has subsided. After examining the IMF condition over the period of 4–8 April shown in Figure 13a, we have chosen 00:00 UT on 7 April (marked by the second vertical dashed line) as the beginning of thermospheric recovery phase when the solar wind

Figure 11. Similar to Figure 8 but for meridional winds at 200 km.

returned to its nominal condition, and both IMF B_y and B_z became small with a magnitude less than 5 nT. The calculated e -folding recovery time τ for global mean temperature and mass density ranges from 27.7 h to 31.5 h, with a median value of 29.5 h. However, for global mean mass density at 150 km, the recovery time is over 70 h, despite the quite small ($<20\%$) overall storm-time density change at that altitude. The e -folding recovery time for the orbitally averaged density from the three satellites ranges from 11.0 h to 19.6 h. However, the e -folding time calculated from the corresponding model results is about 6 h longer than the satellite measurements.

The recovery time for the orbitally averaged mass density from spacecraft observations and model results is shorter than the recovery time for the globally mean mass density from the model. The reason for such discrepancy may lie in the fact that the orbitally averaged mass density does not fully captured the storm-induced thermospheric changes. This can be understood by examining Figures 5–7. For example, the neutral mass density measured by GOCE and also model outputs sampled along the satellite track showed that the density enhancement during the recovery phase after 00 UT on 7 April was limited within the latitude range between 50°N and 20°S as shown in Figure 5 (top and middle rows). However, the modeled neutral density at fixed altitude showed density enhancement at all latitudes from the North Pole to the South Pole. Similarly, the density enhancements seen by GRACE and also by sampling model outputs along the satellite track were both limited to the Northern Hemisphere during the recovery phase so that the density enhancement in the Southern Hemisphere was absent in the top two rows in Figure 7.

4. Summary and Discussion

The combined observational and numerical modeling results presented in this paper offer a rather complete picture of the large-scale ionospheric and thermospheric response to the 5 April 2010 geomagnetic storm. The qualitative agreement between our simulation results and satellite observations indicates that the TIMEGCM, when properly driven by realistic storm-time magnetospheric forcing, is able to reproduce many observed features in the neutral winds and neutral mass density. The main findings are summarized below.

The storm was initially triggered by the compressed solar wind sheath region preceding a fast-moving ICME, which arrived at the Earth on 5 April 2010. The highly fluctuating IMF within the sheath caused dynamic changes in the high-latitude ionosphere as shown in the AMPERE magnetic field perturbations and also by the ionospheric convection patterns derived from AMIE. The TIMEGCM simulations showed that the winds did not exactly follow the $\mathbf{E} \times \mathbf{B}$ drifts; the wind response over the polar cap was more subtle when compared to the ion drifts under rapidly evolving IMF conditions. However, the high-latitude wind patterns did go through some large-scale reconfiguration as illustrated in Figure 4. The cross-track winds measured by GOCE

Figure 12. Similar to Figure 8 but for meridional winds at 100 km.

offered a first glimpse of the model performance in simulating the storm-time neutral wind distributions. A good qualitative agreement was found between the GOCE measurements and the TIMEGCM simulations. Particularly, the model was able to reproduce the similar cross-track wind strength and the wind reversal or reduction near the dawn and dusk polar cap boundary. Storm-related changes in the winds became more prominent in the midlatitude and low-latitude regions, where the winds were perturbed by TADs. Though wind speed is much smaller at midlatitudes and low latitudes than at high latitudes, the winds swing between eastward and westward and also flip between northward and southward as TADs pass through the local regions.

Numerical simulation results presented in this paper clearly demonstrated that, while TADs are global phenomena, they are highly nonuniform, possessing strong latitudinal and longitudinal variations as well as altitude dependence. At 400 km, the TAD phase speed varied from ~ 640 m/s to 780 m/s, depending on local time and hemisphere. The strength of TADs diminished with decreasing altitude. The TAD phase speed decreased with decreasing altitude, roughly by $\sim 5\%$ for an altitude drop of 100 km. In the lower thermosphere near 100 km, TADs became nearly unrecognizable; instead, they are overpowered by tides that propagate into the lower thermosphere from the lower atmosphere.

The complex spatial and temporal varying nature makes the detection of TADs sometimes difficult. This event was a rare case when comprehensive thermospheric observations were made concurrently by three spacecraft: GOCE, CHAMP, and GRACE. Even with this fortuitous satellite coverage, thermospheric disturbances such as TADs were not adequately captured due to their localized and temporally evolving nature. This clearly speaks to the fact that the current observational system is insufficient in monitoring the near-Earth space environment. Global numerical models serve as an important asset to fill the data gaps and also provide a global context to help interpret localized observations.

During geomagnetic storms while the thermosphere adjusts itself to the varying energetic input from the ionosphere and magnetosphere through dynamical/electrodynamical interaction, it also feeds back and regulates this interaction. One outstanding question is: how effective is this feedback interaction on the coupled system? By comparing the variations of global mean temperature and neutral mass density with the globally integrated Joule heating, it was found that the initial thermospheric response to the storm was nearly simultaneous at all altitudes above about 150 km. However, the peak of thermospheric storm enhancement lagged Joule heating by 2–3 h. After storm-time energy input had subsided, based on the coupled AMIE-TIMEGCM simulations, the thermosphere took about 29.5 h to recover in a global mean sense. The orbitally averaged neutral density from GOCE, CHAMP, and GRACE indicated a recovery time between 11.0 to 19.6 h, whereas the similar calculation based on simulation results sampled along the satellite tracks.

Figure 13. Distributions of (a) the IMF B_y and B_z components, (b) the reserved Dst (black) and globally integrated Joule heating rate (red) derived from AMIE, with the 5 min Joule heating shown by the red dotted line and the 3 h running average by the solid red line, (c) percent change of TIMEGCM global mean temperature at selected heights, (d) percent change of TIMEGCM global mean mass density at selected heights, and (e) percent change of orbitally averaged mass density along the GOCE, CHAMP, and GRACE trajectories, with observations shown by solid lines and model results by dashed lines. The maximum cross-correlation coefficient and the corresponding lag time between the reversed Dst and the 3 h running average Joule heating are listed in Figure 13b. τ is the e -folding recovery time at selected height. Note that the value outside (inside) of the parentheses in Figure 13e is from satellite data (the model simulation). The first vertical dashed line marks the onset of the thermospheric response to the storm, and the second vertical dashed line indicates the beginning of the thermospheric recovery phase.

was 6 h longer, varying from 17.2 to 25.6 h. The e -folding recovery time in this case was much longer than the 4–6 h recovery time reported in *Lei et al.* [2011] for the Halloween storm of October 2003 based on CHAMP and GRACE data. The difference between these two cases implies that thermospheric recovery time may depend on the strength of a given storm. Furthermore, the October 2003 storm was near the peak of solar cycle 23, whereas the April 2010 storm was at the rising phase of a much weaker cycle 24. It is possible that the preconditioning of the thermosphere may affect its overall recovery. Thus, more storm studies are needed to understand thermospheric recovery time as functions of storm intensity, solar cycle, and seasons. *Lei et al.* [2011] also noted a longer-recovery time from the Thermosphere-Ionosphere-Electrodynamics General Circulation Model (TIEGCM) simulations than the satellite observations, similar to what we found in this case. Though no definitive answer had been found why the model predicted a longer-recovery phase for the thermosphere, one possible cause may be that the NO cooling rate was underestimated in the model. This calls for a further evaluation of the rate coefficients currently being implemented in the model [e.g., *Lu et al.*, 2010]. Furthermore, the mean energy and energy flux of precipitating electrons are obtained from the AMIE procedure based on various data sets and used as part of the upper boundary input to the TIMEGCM. A simple Maxwellian distribution is assumed to calculate the ionization rate in the TIMEGCM, which is admittedly

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Figure 14. Profile of global mean vertical eddy diffusion coefficient in the standard TIMEGCM setup. The dashed line represents eddy diffusion due to gravity wave breaking, and the dotted line is the background eddy diffusion. The total vertical eddy diffusion is shown by the solid line, which has a maximum value of $225 \text{ m}^2 \text{ s}^{-1}$ near 100 km.

altitude, the reduction in overall O number density results in an altitude dependence of the percentage change in neutral density that increases with altitude. This explains why the difference was negligible at the GOCE orbital altitude of ~ 270 km but became a factor of 2 at the GRACE orbital altitude of ~ 470 km. Qian *et al.* [2009] found that, by increasing the eddy diffusion coefficient by a factor 2 at the lower boundary of 97 km in the TIEGCM, the neutral density at 400 km could be reduced by 40%. For our case presented here, it is possible that the eddy diffusion coefficient as currently implemented in the TIMEGCM might be too high. Figure 14 shows the profile of the global mean vertical eddy diffusion coefficient, which includes eddy diffusion due to gravity wave breaking (dashed line) and background eddy diffusion (dotted line) as described by Liu and Roble [2002]. The calculation of eddy diffusion in the TIMEGCM is subject to large uncertainties associated with the parameterization of gravity waves that cannot be directly resolved by the model and the assumed values for the Prandtl number. A recent study by Garcia *et al.* [2014] found that the standard eddy diffusion value in the TIMEGCM is too large when comparing the calculated CO_2 with observations, and they suggested a reduction of the eddy diffusion coefficient by a factor of 5. We will carefully examine the effect of eddy diffusion in a follow-on study by modifying the parameterization of the relevant eddy coefficients in the TIMEGCM in a similar fashion as described in Qian *et al.* [2009].

Data-model intercomparison is absolutely crucial for model validation. While the TIMEGCM was able to capture the salient features of the neutral winds and mass density throughout the storm, there were qualitative differences between the satellite in situ measurements and the simulation results. This study has pointed out some specific weaknesses in the TIMEGCM that require further improvements in order to meet the challenges of accurate space weather prediction.

References

- Allen, J. (2010), The Galaxy 15 Anomaly: Another satellite in the wrong place at a critical time, *Space Weather*, 8, S06008, doi:10.1029/2010SW000588.
- Anderson, B. J., H. Korth, C. L. Waters, D. L. Green, V. G. Merkin, R. J. Barnes, and L. P. Dyrud (2014), Development of large-scale Birkeland currents determined from the Active Magnetosphere and Planetary Electrodynamics Response Experiment, *Geophys. Res. Lett.*, 41, 3017–3025, doi:10.1002/2014GL059941.
- Balthazor, R. L., and R. J. Moffett (1999), Morphology of large-scale traveling atmospheric disturbances in the polar thermosphere, *J. Geophys. Res.*, 104(A1), 15–24, doi:10.1029/1998JA900039.
- Bruinsma, S. L., and J. M. Forbes (2009), Properties of traveling atmospheric disturbances (TADs) inferred from CHAMP accelerometer observations, *Adv. Space Res.*, 43, 369–376, doi:10.1016/j.asr.2008.10.031.
- Bruinsma, S. L., E. Doornbos, and B. R. Bowman (2014), Validation of GOCE densities and thermosphere model evaluation, *Adv. Space Res.*, 54, 576–585, doi:10.1016/j.asr.2014.04.008.

inaccurate. Precipitating protons are not included in the present study due to the lack of adequate global measurements. Therefore, the production of NO in the model is very likely underestimated.

Comparison with the satellite measurements revealed that the model underestimates neutral mass density in the upper thermosphere above 300 km. As noted by Qian *et al.* [2009], the upper thermospheric mass density is strongly modulated by eddy mixing in the mesosphere and lower thermosphere due to gravity wave breaking. Eddy diffusion in the lower thermosphere reduces atomic oxygen O through downward transport. The reduction of O number density then propagates to the upper thermosphere through molecular diffusion. Since the contribution of O to the total neutral density increases with

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- Bruinsma, S., J. M. Forbes, R. S. Nerem, and X. Zhang (2006), Thermosphere density response to the 20–21 November 2003 solar and geomagnetic storm from CHAMP and GRACE accelerometer data, *J. Geophys. Res.*, **111**, A06303, doi:10.1029/2005JA011284.
- Clilverd, M. A., C. J. Rodger, D. Danskin, M. E. Usanova, T. Raita, T. Ulich, and E. L. Spanswick (2012), Energetic particle injection, acceleration, and loss during the geomagnetic disturbances which upset Galaxy 15, *J. Geophys. Res.*, **117**, A12213, doi:10.1029/2012JA018175.
- Connors, M., C. T. Russell, and V. Angelopoulos (2011), Magnetic flux transfer in the 5 April 2010 Galaxy 15 substorm: An unprecedented observation, *Ann. Geophys.*, **29**, 619–622, doi:10.5194/angeo-29-619-2011.
- Denig, W. F., J. C. Green, D. C. Wilkinson, J. V. Rodriguez, H. J. Singer, P. T. Loto'Aniu, D. A. Biesecker, and W. Murtagh (2010), Space Weather Conditions at the Time of the Galaxy 15 Spacecraft Anomaly, AGU, Fall Meeting 2010, abstract #SH31D-03.
- Doornbos, E., J. van den IJssel, H. Lühr, M. Förster, and G. Koppenwallner (2010), Neutral density and crosswind determination from arbitrarily oriented multi-axis accelerometers on satellites, *J. Spacecraft Rockets*, **47**(4), 580–589, doi:10.2514/1.48114.
- Doornbos, E., S. Bruinsma, B. Fritsche, G. Koppenwallner, P. Visser, J. van den IJssel, and J. de Teixeira de Encarnação (2014), ESA Contract 4000102847/NL/EL GOCE+ Theme 3: Air density and wind retrieval using GOCE data—Final report, TU Delft, Netherlands.
- Floberghagen, R., M. Fehringer, D. Lamarre, D. Muzi, B. Frommknecht, C. Steiger, J. Pineiro, and A. da Costa (2011), Mission design, operation and exploitation of the gravity field and steady-state ocean circulation explorer mission, *J. Geodyn.*, **85**, 749–758, doi:10.1007/s00190-011-0498-3.
- Forbes, J. M., G. Lu, S. Bruinsma, S. Nerem, and X. Zhang (2005), Thermosphere density variations due to the 15–24 April 2002 solar events from CHAMP/STAR accelerometer measurements, *J. Geophys. Res.*, **110**, A12527, doi:10.1029/2004JA010856.
- Fuller-Rowell, T. J., M. V. Codrescu, R. J. Moffett, and S. Quegan (1994), Response of the thermosphere and ionosphere to geomagnetic storms, *J. Geophys. Res.*, **99**(A3), 3893–3914, doi:10.1029/93JA02015.
- Fuller-Rowell, T. J., M. V. Codrescu, R. G. Roble, and A. D. Richmond (1997), How does the thermosphere and ionosphere react to a geomagnetic storm?, in *Magnetic Storms*, Geophys. Monograph Series, vol. 98, edited by B. T. Tsurutani et al., pp. 203–225, AGU, Washington, D. C.
- Garcia, R. R., M. López-Puertas, B. Funke, D. R. Marsh, D. E. Kinnison, A. K. Smith, and F. González-Galindo (2014), On the distribution of CO₂ and CO in the mesosphere and lower thermosphere, *J. Geophys. Res. Atmos.*, **119**, 5700–5718, doi:10.1002/2013JD021208.
- Gonzalez, W. D., J. A. Joselyn, Y. Kamide, H. W. Kroehl, G. Rostoker, B. T. Tsurutani, and V. M. Vasyliunas (1994), What is a geomagnetic storm?, *J. Geophys. Res.*, **99**(A4), 5771–5792, doi:10.1029/93JA02867.
- Hagan, M. E., and J. M. Forbes (2002), Migrating and nonmigrating diurnal tides in the middle and upper atmosphere excited by tropospheric latent heat release, *J. Geophys. Res.*, **107**(D24), 4754, doi:10.1029/2001JD001236.
- Hagan, M. E., and J. M. Forbes (2003), Migrating and nonmigrating semidiurnal tides in the upper atmosphere excited by tropospheric latent heat release, *J. Geophys. Res.*, **108**(A2), 1062, doi:10.1029/2002JA009466.
- Kinrade, J., C. N. Mitchell, P. Yin, N. Smith, M. J. Jarvis, D. J. Maxfield, M. C. Rose, G. S. Bust, and A. T. Weatherwax (2012), Ionospheric scintillation over Antarctica during the storm of 5–6 April 2010, *J. Geophys. Res.*, **117**, A05304, doi:10.1029/2011JA017073.
- Lei, J., J. P. Thayer, G. Lu, A. G. Burns, W. Wang, E. K. Sutton, and B. A. Emery (2011), Rapid recovery of thermosphere density during the October 2003 geomagnetic storms, *J. Geophys. Res.*, **116**, A03306, doi:10.1029/2010JA016164.
- Liu, H., H. Lühr, S. Watanabe, W. Köhler, V. Henize, and P. Visser (2006), Zonal winds in the equatorial upper thermosphere: Decomposing the solar flux, geomagnetic activity, and seasonal dependencies, *J. Geophys. Res.*, **111**, A07307, doi:10.1029/2005JA011415.
- Liu, H.-L., and R. G. Roble (2002), A study of a self-generated stratospheric sudden warming and its mesospheric-lower thermospheric impacts using the coupled TIME-GCM/CCM3, *J. Geophys. Res.*, **107**(D23), 4695, doi:10.1029/2001JD001533.
- Lu, G. (2006), High-speed streams, coronal mass ejections, and interplanetary shocks: A comparative study of geoeffectiveness, in *Recurrent Magnetic Storms: Corotating Solar Wind Streams*, AGU Geophys. Monogr. Ser., vol. 167, edited by B. Tsurutani et al., pp. 97–111, AGU, Washington, D. C.
- Lu, G., et al. (1998), Global energy deposition during the January 1997 magnetic cloud event, *J. Geophys. Res.*, **103**(A6), 11,685–11,694, doi:10.1029/98JA00897.
- Lu, G., A. D. Richmond, J. M. Ruohoniemi, R. A. Greenwald, M. Hairston, F. J. Rich, and D. S. Evans (2001a), An investigation of the influence of data and model inputs on assimilative mapping of ionospheric electrodynamics, *J. Geophys. Res.*, **106**, 417–433, doi:10.1029/2000JA000606.
- Lu, G., A. D. Richmond, R. G. Roble, and B. A. Emery (2001b), Coexistence of ionospheric positive and negative storm phases under northern winter conditions: A case study, *J. Geophys. Res.*, **106**, 24,493–24,504, doi:10.1029/2001JA000003.
- Lu, G., L. P. Goncharenko, A. D. Richmond, R. G. Roble, and N. Aponte (2008), A dayside ionospheric positive storm phase driven by neutral winds, *J. Geophys. Res.*, **113**, A08304, doi:10.1029/2007JA012895.
- Lu, G., M. G. Mlynarczyk, L. A. Hunt, T. N. Woods, and R. G. Roble (2010), On the relationship of Joule heating and nitric oxide radiative cooling in the thermosphere, *J. Geophys. Res.*, **115**, A05306, doi:10.1029/2009JA014662.
- Lu, G., et al. (2011), Reversed two-cell convection in the Northern and Southern Hemispheres during northward interplanetary magnetic field, *J. Geophys. Res.*, **116**, A12237, doi:10.1029/2011JA017043.
- McComas, D. J., N. Buzulukova, M. G. Connors, M. A. Dayeh, J. Goldstein, H. O. Funsten, S. Fuselier, N. A. Schwadron, and P. Valek (2012), Two wide-angle imaging neutral-atom spectrometers and interstellar boundary explorer energetic neutral atom imaging of the 5 April 2010 substorm, *J. Geophys. Res.*, **117**, A03225, doi:10.1029/2011JA017273.
- Möstl, C., M. Temmer, T. Rollett, C. J. Farrugia, Y. Liu, A. M. Veronig, M. Leitner, A. B. Galvin, and H. K. Biernat (2010), STEREO and Wind observations of a fast ICME flank triggering a prolonged geomagnetic storm on 5–7 April 2010, *Geophys. Res. Lett.*, **37**, L24103, doi:10.1029/2010GL045175.
- Prikryl, P., et al. (2011), Interhemispheric comparison of GPS phase scintillation at high latitudes during the magnetic-cloud-induced geomagnetic storm of 5–7 April 2010, *Ann. Geophys.*, **29**, 2287–2304, doi:10.5194/angeo-29-2287-2011.
- Prölss, G. W. (1993), On explaining the local time variation of ionospheric storm effects, *Ann. Geophys.*, **11**, 1–9.
- Qian, L., S. C. Solomon, and T. J. Kane (2009), Seasonal variation of thermospheric density and composition, *J. Geophys. Res.*, **114**, A01312, doi:10.1029/2008JA013643.
- Richmond, A. D., and Y. Kamide (1988), Mapping electrodynamic features of the high-latitude ionosphere from localized observations: Technique, *J. Geophys. Res.*, **93**(A6), 5741–5759, doi:10.1029/JA093iA06p05741.
- Richmond, A. D., and S. Matsushita (1975), Thermospheric response to a magnetic substorm, *J. Geophys. Res.*, **80**(19), 2839–2850, doi:10.1029/JA080i019p02839.
- Rishbeth, H. (1989), F-region storms and thermospheric circulation, in *Electromagnetic Coupling in the Polar Cleft and Caps*, edited by P. E. Sandholt and A. England, pp. 393–406, Kluwer Acad. Norwell, Mass.
- Roble, R. G. (1995), Energetics of the mesosphere and thermosphere, in *The Upper Mesosphere and Lower Thermosphere: A Review of Experiment and Theory*, Geophys. Monogr. Ser., vol. 87, pp. 1–21, AGU, Washington D. C.

- Roble, R. G., and E. C. Ridley (1994), A thermosphere-ionosphere-mesosphere-electrodynamics general circulation model (time-GCM): Equinox solar cycle minimum simulations (30–500 km), *Geophys. Res. Lett.*, *21*, 417–420, doi:10.1029/93GL03391.
- Sieradzki, R., I. Chemiak, and A. Krankowski (2013), Near-real time monitoring of the TEC fluctuations over the Northern Hemisphere using GNSS permanent networks, *Adv. Space Res.*, *52*(3), 391–402, doi:10.1016/j.asr.2013.03.036.
- Sutton, E. K., J. M. Forbes, and R. S. Nerem (2005), Global thermospheric neutral density and wind response to the severe 2003 geomagnetic storms from CHAMP accelerometer data, *J. Geophys. Res.*, *110*, A09S40, doi:10.1029/2004JA010985.
- Wilder, F. D., G. Crowley, S. Eriksson, P. T. Newell, and M. R. Hairston (2012), Ionospheric Joule heating, fast flow channels, and magnetic field line topology for IMF B_y -dominant conditions: Observations and comparisons with predicted reconnection jet speeds, *J. Geophys. Res.*, *117*, A11311, doi:10.1029/2012JA017914.