Wave signatures in the midlatitude ionosphere during a sudden stratospheric warming of January 2010

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1. Introduction

The pronounced day-to-day variability in ionospheric parameters remains an area of active research, as physical causes and drivers of this variability are not well understood. Many prior studies focus on sources of variability in the F region peak electron density, mostly because of data availability from ionosondes. Plasma temperatures, and in particular ion temperature \(T_i\), are studied less extensively, though they remain an important factor characterizing the thermal properties of the ionosphere.

The general behavior of \(T_i\) is governed by the 11 yr solar cycle, seasonal cycle, and diurnal cycle associated with solar heating, and it strongly varies depending on local time, altitude, and latitude. A large portion of day-to-day variability in plasma temperature is controlled by short-term variations in the solar and geomagnetic activities. Numerous studies have summarized the current understanding of plasma temperature behavior in empirical models based either on satellite [Triskova et al., 1996; Truhlik et al., 2000; Bilitza, 2001] or ground-based data [e.g. Holt et al., 2002; Zhang and Holt, 2007]. Such models usually describe variations in ionospheric parameters due to solar and geomagnetic activities as a function of the previous day's 10.7 cm solar flux \(F_{10.7}\) and a 3 h \(Kp\) index. Deviations of actual data from models are thought to result from a combination of contributions from several drivers, such as uncertainties in specification of solar EUV flux, high-latitude forcing, and forcing of a meteorological nature (gravity waves, tides, planetary waves). Zhang and Holt [2008a] reported that for the midlatitude ionosphere under geomagnetically quiet conditions \((ap=0–30)\) and winter solstice, the standard deviation of \(T_i\) from climatological values is between 130 and 170 K, with higher standard deviation values observed at higher altitudes, and 78% of the data falling within \pm\ one standard deviation. The background variability, which is independent from solar and geomagnetic activities, has a clear seasonal and height variation, with minimums of \(-100\) K observed at winter and summer solstices at 230–250 km. As ions and neutrals

[1] This paper presents a case study of the day-to-day variability in the midlatitude upper atmospheric ion temperature (~200–400 km) with a focus on variability resulting from meteorological forcing. The data are obtained by the Millstone Hill incoherent scatter radar (42.6°N, 288.5°E) on 18–31 January 2010, in coincidence with a major sudden stratospheric warming. We elucidate oscillations in ion temperature with both tidal periods (~8 h and ~12 h) and non-tidal periods (>24 h) by analyzing residuals between the observed temperatures and those expected from an empirical model. We present the spatial-temporal development of periodicities in ion temperature and discuss to what degree these periodicities might be related to the sudden stratospheric warming event. The spectral location and temporal evolution of periodicities with ~9.9–12.9 h and ~6.2–7.9 h suggest that they are related to the semiannual (12 h) and terdiurnal (8 h) tides that are enhanced during the sudden stratospheric warming. Periodicities with ~3–4 d and ~10–13 d are likely related to Rossby waves with 4 d and 10 d periods, while the strong periodicity observed at 16–17 h could result from the nonlinear interaction of the quasi 2 d wave with the semiannual tide. As planetary waves are not expected to propagate to altitudes of ~200–250 km, these experimental results raise questions about the potential mechanisms of coupling between the lower and upper atmosphere.

are in close thermal contact at altitudes below ~250 km, variability in $T_\parallel$ can serve as a good representation of variability in neutral temperature and provides insights about the coupling between the lower and upper atmospheric regions.

Furthermore, meteorological influences from the lower atmosphere can propagate to the thermosphere and create perturbations in the ionosphere as well. Atmospheric tides generate ionospheric disturbances at harmonics of the 24 h rotation period of Earth, while planetary waves at 2, 5, 10, and 16 d oscillations are known to populate the lower thermosphere/ionosphere. The existence of oscillations in Earth’s ionosphere at periods of 2, 5, 6–7, 9–10, and 12–18 d is well established [e.g., Altadill et al., 2003; Altadill and Apostolov, 2003; Forbes and Leveroni, 1992; Forbes et al., 1997; Laštovička et al., 2003]. These are referred to as oscillations at planetary wave (PW) periods since they correspond to waves in the middle atmosphere that are characteristic of normal modes or resonant oscillations in the case of 2, 5, 9–10, and 12–18 d periods [Salby, 1984], or of maximum growth rates of unstable regions [e.g., Meyer and Forbes, 1997a; Liu et al., 2004], and/or wave-wave interactions [Pogoreltsev et al., 2002] in the case of the 6–7 d wave. Thus, planetary waves propagating from the lower atmosphere are an alternative source of variability in the ionosphere. Theoretical and modeling studies demonstrate that planetary waves are not capable of propagating above about 100–110 km [Pogoreltsev et al., 2007]. However, $F$ region ionospheric observations have demonstrated periodic oscillations that have been related to planetary wave periods. Although several mechanisms have been proposed (see review by Forbes [1996]), the manner in which planetary waves may drive ionospheric variations still remains largely unknown, and therefore, more indirect scenarios, such as modulation of dynamo-generated electric fields that map into the $F$ region, appear to be required.

Sudden stratospheric warming (SSW) is a large-scale phenomenon in the winter polar middle atmosphere that is characterized by a rapid increase in stratospheric temperature and dynamics [O’Neill, 2003] and is forced by an anomalous increase in quasi-stationary planetary wave activity. SSW events greatly influence chemical composition, electrodynamics, and mixing and transport processes of the atmosphere and ionosphere. This phenomenon has recently attracted significant attention as an innovative approach for studying coupling between the lower and upper thermosphere in a planned and deliberate way, as SSW can be predicted up to 8–10 d in advance. Experimental studies suggest that SSW couples all atmospheric layers from the ground to the thermosphere and from the poles to the equator and the opposite hemisphere [Chau et al., 2009; Fejer et al., 2010, 2011; Goncharenko et al., 2010a, 2010b; Liu et al., 2011a, 2011b; Pancheva and Mukhtarov, 2011; and review by Chau et al., 2012]. The resulting disturbances in total electron content are as large as a 50–150% variation from a mean state [Goncharenko et al., 2010a]. These disturbances are interpreted in terms of solar and lunar tides that are amplified during the SSW events and affect the low-latitude $F$ region through the $E$ region dynamo mechanism [Pedatella and Forbes, 2010; Fejer et al., 2010; Liu et al., 2010; Fuller-Rowell et al., 2011].

2. Data and Methods

For this work, we use measurements of ionospheric parameters obtained by the Millstone Hill ISR UHF zenith antenna from 18–31 January 2010 operating in single pulse mode (~18 km altitude resolution). For comparison, we also use data at the same range of altitudes as the 2010 SSW campaign: 20–23 January 2007; 17–21 and 25–28 January 2008, before and after the peak of the SSW event, as well as data from 4, 5, and 11–14 February 2009, e.g., after the peak of the SSW event. The data run continuously throughout all of the observing periods and provided accurate estimates of ionospheric parameters in the $F$ region.

Our approach for this study consists of filtering the data, constructing an empirical model to characterize the $T_\parallel$ behavior at the $F$ region for low solar and geomagnetic activities, identifying periodicities in the data through spectral analysis, and reconstructing the oscillations according to the main periodicities detected. The January 2007 and 2008 and February 2009 data were introduced in...
this work for the empirical model development in order to increase the number of available points in the solar and geomagnetic dependencies in the data. In 2009, a major SSW event occurred with its mature stage around 24 January. Therefore, the model uses data from 2009 that are spaced 10+ d after the peak of the event. Detailed information about the January 2007 and 2008 stratospheric and geophysical conditions may be found in Goncharenko and Zhang [2008].

Figure 1 summarizes the stratospheric and geophysical conditions for the 2010 winter campaign period presented in this paper. Figures 1a and 1b show National Centers for Environmental Prediction stratospheric temperatures at 10 hPa (~30 km) for 90°N and the zonally averaged temperatures for 60–90°N, respectively (black dots). The 30 yr median temperatures (light gray circles) for the January 2010 period are also shown for comparison. The data for 90°N present lower temperatures before the SSW event (by ~16 K, black dots) when compared with the historical mean. Between 19 and 22 January, there is an increase in the stratospheric temperatures of ~9.5 K per day at 90°N, indicative of the sudden stratospheric warming event (maximum value ~235.6 K on 22 January, which is ~25 K greater than the historical mean and ~40 K greater than the temperatures before the beginning of the SSW). A similar behavior is observed in the zonally averaged temperatures for 60–90°N, but the increase in temperature that begins on 19 January presents a smaller rate of temperature change (~1.82 K per day) when compared with temperatures at 90°N. This temperature increase encompasses the whole period of study. The stratospheric circulation is characterized by the zonal mean zonal wind at 60°N at the same pressure level (10 hPa) and is presented in Figure 1c. Before the SSW period, the zonal wind was ~45 ms per day, larger than the 30 yr average. However, during the SSW event, the zonal neutral wind component decreases to almost zero (less eastward) from 19 January to 27 January, with a rate of ~4.5 ms per day. Finally, based on the 30 yr median values, the temperatures for 90°N and the zonally averaged temperatures for 60–90°N returned to normal conditions on 23 and 21 February, respectively, while the zonal mean zonal wind at 60°N returned to its mean state on 2 March (not shown in this paper). This SSW event was driven by amplification in planetary wave 1 activity (Figure 1d) that strongly increased by mid-January and returned to the average level of activity by the end of January. Planetary wave 2 (Figure 1e) remained below the average level of winter-time activity throughout the entire month of January 2010. The SSW event in this study is during a period of low-to-disturbed geomagnetic activity fluctuations (Kp varies mostly between 0 and 2 and briefly increases to 4 on 20 January, as seen in Figure 1g) and during low solar activity (~80 solar flux unit (sfu); 1 sfu = 1022 W m–2 Hz–1) shown in Figure 1f.

2.1. Data Filtering and Processing Methodology

We limit this study to Ti variability in the altitude range of 195.5–393.3 km because of data gaps that exist outside this altitude range. We note that the type of spectral analysis applied to this study cannot be performed at altitudes with discontinuous data. The standard ISR data analysis procedure outputs missing values when the electron density (Ne) is low, due to the low signal-to-noise ratio (SNR). In addition to the prevalence of missing values below 200 km and above 400 km during the nighttime and for low solar activity, the errors in Ti are too large to be considered reliable data points.

To account for any missing data that occurred in the altitude region of study, which were relatively infrequent and isolated to a narrow altitude range, we applied a linear interpolation between the values before and after the data gap. The effect of this interpolation on the spectral analysis depends on the maximum time length of the data gap. That is, any periodicities less than the maximum time gap will be altered by the interpolation, while periodicities greater
than this time length will remain the same. For January 2007, 2008, and 2010 and February 2009, there existed no more than three consecutive gaps in the data points. This resulted in a maximum time gap in the data of 12.6 min for January 2007 and 2010 and 4 and 5 February 2009, and 48.6 min for January 2008 and 11–14 February 2009. Since we are interested in periodicities at much lower frequencies (greater than four hour time lengths), the interpolation does not affect the results of our wavelet analysis.

[13] We applied a filtering algorithm to remove noise and outliers in the data. Figure 2 summarizes the filtering method. The filtering process pertained only to the $T_i$ data with high frequency periodicities, or noise (<4.48 h for January 2007 and 2010 and 4 and 5 February 2009; and <4.32 h for January 2008 and 11–14 February 2009). The $T_i$ signal was decomposed using the Daubechies wavelet function and the noise was isolated. A standard deviation was computed for the difference between the original data and the data with the noise removed (blue and red lines of Figure 2a) that located outliers and replaced them with the noise-free data. This filtering process was repeated until the standard deviation became asymptotic. This filter window was applied with respect to time at each altitude, and it ensured that the $T_i$ data approached its ideal noise-free state. The $T_i$ signal approaches its ideal state after approximately 10 filter cycles (see Figure 2b, where $N$ is the number of cycles). The standard $z$ score used for January 2007, 2008, 2009, and 2010 was 1.29, which means that approximately 20% of the $T_i$ values were subject to replacement in the first cycle (assuming a natural Gaussian distribution) by the data without noise, and in each subsequent cycle of the filtering process, this percentage decreases until a condition where none of the data are replaced is reached (as seen in Figure 2c).

[14] Residuals were determined from the filtered $T_i$ values minus the values obtained by an empirically derived ion temperature “quiet time” model, which will be described in the subsequent section. Spectral analysis was performed on the hourly averaged $T_i$ residuals using the Morlet wavelet as the mother function.

2.2. Empirical $T_i$ Model

[15] Ion temperatures exhibit substantial spatial and temporal variation over a wide range of scales with different dependencies. Two important geophysical parameters influencing $T_i$ are solar and geomagnetic activities [Forbes et al., 2000]. Each one of these parameters influences the $T_i$ behavior in different ways depending on local time for the same season. For this reason, it is necessary to know the individual contribution of each parameter to the modulation of $T_i$ in order to better understand their response to an SSW event. For the prediction-estimation of the $T_i$ behavior in our model, we take into account the contribution of solar and geomagnetic activities as a function of local time, ignoring the seasonal dependence since we are working only with data between 17 January and 14 February. It is well known that the response of the middle- and low-latitude thermosphere to magnetic activity may be delayed by several hours after an increase of energy deposition into the high-latitude ionosphere [Duboin and Lafleur, 1992; Fejer et al., 2002]. For the Millstone Hill location, the time

![Figure 2](image-url)
The main source of the energy deposition from solar irradiance into the atmosphere is the UV-EUV emission lines. The solar UV-EUV spectrum, which is composed of emission lines superimposed on a relatively weak continuum, has origins in the Sun’s chromosphere, transition region, and corona. These regions of the solar atmosphere respond dramatically to magnetic activity in the Sun, causing the UV-EUV irradiances to vary significantly throughout the solar cycle [Lean, 1990]. The increase of the UV-EUV dispersion with solar activity reflects temporal and spatial structures between source regions of chromospheric and coronal radiation emissions [Liu et al., 2006 after Lean et al., 2001]. Brum et al. [2011] and Brum et al. [2012] have shown that the best description of the UV-EUV is a combination of the daily decimetric solar flux index \( F_{107} \) and one more term \( F_{107a} \), which corresponds to the average of the 81 previous days \( (F_{107} + F_{107a})/2 \). In addition, their works have shown that the UV-EUV emissions tend to increase with \( F_{107} \) until a certain threshold (around 175 sfu). However, for low solar activity, the UV-EUV variations to the \( F_{107} \) can be well represented by a linear function.

In order to reproduce the \( T_i \) behavior due to the solar and geomagnetic variations, we extracted the linear dependence of the \( T_i \) as a function of the previously cited indices. The method applied is based on regression analysis, in which observational data are modeled by a function that is a linear combination of the model parameters, and depends on one or more independent variables (in this case, the combination of the decimetric solar flux \( F_{107} \), and \( ap_0, ap_{-3} \), and \( ap_{-6} \) indices, see Figure 3). We estimated the best fit (in a linear sense) of the data with the independent variables of interest simultaneously. With this method, we obtained the relative contribution of each individual geophysical parameter to the variability of the \( T_i \) signal. The linear reconstruction of the \( T_i \) behavior \( (T_i(p_{alt,t}) \) for a single proxy \( P \) as a function of altitude \( (alt) \) and time \( (t) \) is obtained by

\[
T_i(p_{alt,t}) = a_{P(alt,t)} + b_{P(alt,t)}P \quad \text{for} \quad P = F_{107p}, ap_0, ap_{-3}, ap_{-6} \tag{1}
\]

where \( a_{P(alt,t)} \) and \( b_{P(alt,t)} \) are the real constants of the linear function \( T_i(p_{alt,t}) \) for the dependent variable \( P \). \( a_{P(alt,t)} \) is the value of \( T_i(p_{alt,t}) \) for \( P = 0 \) (\( P \) is a real variable), and \( b_P \) is the \( T_i \) dependence rate (slope) related to the proxy \( P \). Thus, the final reconstructed \( T_i \) is obtained by the sum of all the isolated contributions (see Figure 6).

The time and altitudinal dependencies \((t \text{ and } alt, \text{ respectively})\) of the linear function parameters \((a_{p(alt,t)} \text{ and } b_{p(alt,t)})\) for a single proxy of the model is determined by the Fourier transform reconstruction, given by

\[
[a_{p(alt,t)} \text{ or } b_{p(alt,t)}] = AO_{p(alt)} + \sum_{m=1}^{\infty} \left( A_{p(alt)} \cos(2\pi ft) + B_{p(alt)} \cos(2\pi ft) \right) \tag{2}
\]

where \( AO_{p(alt)} \) is the daily average for a determinate coefficient of the linear function, \( A_{p(alt)} \) and \( B_{p(alt)} \) are the coefficients associated with the harmonic \( m \), and \( f \) represents the fundamental frequency of the signal \( a_{p(alt,t)} \) or \( b_{p(alt,t)} \) given by \( f = m/24 \).

**Figure 3.** Example of how the linear function coefficients of the empirical model were taken. (a) Top block, the scatter plots of the residual of \( T_i \) versus the combination of the decimetric solar flux \( (F_{107p}) \). (b) Bottom block, the dependence of the residual of \( T_i \) with the variation of \( ap_0, ap_{-3} \), and \( ap_{-6} \) (from the top to the bottom) indices for six ranges of time at 285.47 km of altitude. The continuous line in each block represents the best linear fit to the parameters. In order to use the same vertical scale for the solar flux dependence, the \( T_i \) values were normalized to a zero at \( F_{107} = 75 \text{ sfu} \). The values presented in each block of panels related with the geomagnetic dependence are the data minus the dependence of \( T_i \) to the solar activity.
Finally, the altitudinal dependence of the model is obtained by the altitude neighborhood, and it is applied to the Fourier coefficients (those presented in equation (2)) by polynomial fits (following the procedure applied by Garzón et al., 2011). The frequencies included in the model are those related to diurnal (24 h, \( m = 1 \)), semidiurnal (12 h, \( m = 2 \)), terdiurnal (8 h, \( m = 3 \)), and quarter-diurnal (6 h, \( m = 4 \)) periodicities. As an example, Figure 4 displays the \( \beta_{F107P} \) Fourier parameters (\( A_0, A_m, B_m \), and \( B_m \), for \( P = F107P \)) as a function of altitude and the associated best polynomial fits (continuous lines).

The justification for developing a new model for this study, as opposed to utilizing previously established ionospheric models, is supported by Figure 5. In Figure 5, the developed Millstone Hill model (MHm) is compared with the International Reference Ionosphere (IRI) model [Bilitza, 1990, 2001, 2003; Bilitza et al., 2004] and the US Naval Research Laboratory Mass and Spectrometer Incoherent Scatter Radar (NRLMSIS) model [Hedin, 1987; Picone et al., 2002], along with the Incoherent Scatter Radar Ionospheric Model (ISRIM) [Zhang and Holt, 2007]. In addition, the \( T_i \) data averages for the periods outside of the SSW are plotted with their respective dispersions (same data used for the model development, light gray circles; geophysical condition: \( F107P = -75 \) sfu and \( ap \leq 7 \) nT). The IRI and ISRIM models provide \( T_i \) while NRLMSIS outputs only neutral temperature. However, for altitudes below \( \sim 250-400 \) km, the ion temperature is expected to follow the neutral temperature for the midlatitude region [Schunk and Nagy, 2009]. The IRI temperatures are consistently warmer than MHm for altitudes above \( \sim 235 \) km, while NRLMSIS predictions are always lower than MHm. Due to the historical deep solar minimum conditions experienced in 2010, it is expected that IRI would overestimate ion temperatures. Also, despite the Schunk and Nagy [2009] considerations, the NRLMSIS predictions show that for the altitudinal range of study, there exists a “detachment” between the ion and neutral temperatures resulting from a decreasing ion-neutral momentum transfer collision frequency with increasing altitude, which could explain the negative temperature offset observed between NRLMSIS and MHm for the highest altitudes under study. ISRIM is the closest match with the developed model, but there are still significant descrepancies. It is seen that ISRIM also suffers from overprediction of ion temperatures, with the largest differences reported during the daytime. Thus, it was critical to develop a model that included the effects of the deep solar minimum period in order to accurately study the \( T_i \) differences between a non-SSW period and an SSW period.

Finally, Figure 6 gives an overview of the \( T_i \) residual computation. Figures 6a and 6b show the solar and

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**Figure 4.** Altitudinal dependence of the \( \beta_{F107P} \) Fourier parameters. The circles represent the values of the \( A_0, A_m, B_m \), and \( B_m \), corresponding to the harmonic \( m \) (\( m \) varying from 0 to 4), while the continuous lines demonstrate the best polynomial approximation considering the altitude neighborhood dependence.

**Figure 5.** Comparison among the simulations of the developed model (MHm, black lines) and simulations carried out for IRI, IRI, and MSIS (red, blue, and green lines, respectively) under the geophysical conditions given by \( F107P = -75 \) sfu and \( ap \leq 7 \) nT for four different periods. The light gray dots represent the \( T_i \) data mean under the geophysical conditions given by \( F107P = 75 \pm 5 \) sfu and \( ap \leq 7 \) nT, while the error bars indicate a two standard deviation confidence interval.
geomagnetic activities for the January 2010 campaign (blue lines), along with their respective inputs into the empirical model simulation (red lines). Figures 6c and 6d represent the expected variations in $\text{Ti}$ due to the changing solar flux and $\text{ap}$ indices. The light gray line in Figure 6d represents the sum of the geomagnetic contribution to $T_i$, and the red line, including the one standard deviation around the model prediction as a function of day of the year, displays the filtered hourly average of the filtered data determined from subtracting the model from the data.

3. Results and Discussion

3.1. Observed Periodicities

In the previous section, we presented the methodology to obtain the $T_i$ residuals based on the differences between the radar data minus the simulation output obtained under the same geophysical conditions (taking into account the solar and geomagnetic activities). This means that, at least hypothetically, any contributions of solar variation and/or geomagnetic fluctuations in the data were extracted, or attenuated, and any strong periodicity detected in the $T_i$ residuals must be related to different modulators.

Figure 7 presents the residuals for all of the altitudes under study (left column of panels) and their respective power spectrums (right column of panels) from 18–31 January 2010. In addition, in each one of these panels, the vertical bars highlight the “starting phase” of the SSW in study, i.e., the period where the stratospheric temperature increase rate was large compared with the rest of the period in study (see Figure 1a). The black contour lines in the power spectrum panels denote the regions with a 95% confidence level, and the white thicker lines represent the cone of influence (COI). The COI is defined as the region of the wavelet spectrum where edge effects become important (beginning and end of the data series). An enhancement in the power spectrums is observed in the “starting phase” of the SSW from ~231 km to 321 km. During this phase, the corresponding residuals in this altitude range show similar features in their variations, namely, a double-humped peak followed by a dip and subsequent increase in ion temperature before leveling off at the end of the phase. The enhancement in power begins to disappear above 321 km.

Figure 8 presents the integrated power spectrum of the $T_i$ residuals seen between the vertical lines of the right
Figure 7. Power spectrum of the $T_i$ residuals. The left panels show the $T_i$ residuals for all of the altitudes in study (~195.5 km to ~393.3 km, altitude increases from the bottom to top panel), while the right panels depict the periodicities observed throughout the campaign period (periods between ~4 and ~75 h). The black contour lines in the power spectrum maps (right panels) denote the regions with a 95% confidence level, and the white thicker lines represent the cone of influence, while the vertical bars highlight the period where the stratospheric temperature increase rate was large compared with the rest of the period in study.

panels of Figure 7, which have been normalized to the maximum value (maximum = 1). The white lines represent the pronounced periodicities observed as a function of period and altitude. Notice that Figure 8 shows a larger range of periodicities than in Figure 7 (the periodicities shown in Figure 7 are smaller than ~75 h). These are the periodicities that reside outside of the COI and may be interpreted as the trends of the SSW “starting phase.” Five dominant periodicities were detected, as seen in Figure 8: ~7 h, ~16 h, ~36 h, ~90 h, and ~257 h. Also, a 13 h oscillation was observed at lower altitudes, below ~230 km.

[25] We applied the same integration procedure over this time interval in order to obtain the power spectrums for the global, equatorial, and auroral indices ($ap$, $Dst$, and $AE$ indices, respectively) and also for the decimetric solar flux ($F_{10.7}$), shown in Figure 9, to identify any geophysical contributions to the periodicities addressed above. For the solar activity, it is noticed that there is no significant periodicity below ~64 h. The global and auroral indices present a 30.3 h periodicity that might be associated to the sum of diurnal and quarter-diurnal periods and seem to start at the beginning of the SSW event. Finally, all of the geomagnetic indices present a periodicity around ~18 h, and these oscillations start almost together with the SSW event. Analysis of geomagnetic indices for a more extended time period, January and February 2010 (not shown here), reveals a periodicity with a spectral peak at ~2.5 d around 19–23 January and a stronger oscillation with a ~10 d period observed from early January to mid-February. As we do not observe the ~2.5 d periodicities in the $T_i$ residuals, this is a good indication that our methodology properly removes geomagnetic influences from the $T_i$ data.

[26] Figures 10 and 11 show the stronger periodicities detected by the spectrum analyses shown in Figures 7 and 8 during the “starting phase” of the SSW event. Figure 10 shows the periodicities that are less than 1 d, and Figure 11 shows those periodicities that are greater than 1 d. These periodicities correspond to periodicities inside and outside of the COI, respectively. The left panels show the residual decompositions for a specific range of periodicities for selected altitudes in study, while the right panels are the reconstruction of the same signals as a function of time and altitude for all altitudes. The reconstructed $T_i$ signals for 7.7–8.3 h, 9.9–12.9 h, and 14.1–18.2 h in the right column of Figure 10 exhibit similar structures. Warming and cooling trends begin before the peak stratospheric temperature increase rate (22 January), and these perturbations are extended to higher altitudes during the “starting phase” of the event. These characteristics are different from the ones seen in the reconstructed $T_i$ signals for longer periodicities (see right panels of Figure 11). The feature of “lifted” trends to higher altitudes that is observed in the $T_i$ residuals in Figure 10 is not observed in Figure 11. This feature could be due to some mechanism on the bottomside that may influence the higher
periodicities (between ~200–260 km and ~200–360 km for 10–13 h and 14–18 h, respectively). As seen in Figure 10, this “lifted” feature apparently occurs in phase with the “starting phase” of the SSW and with the increase in geomagnetic activity. Notice in Figure 7 that the enhancements in power are not seen at altitudes lower than 231 km but are seen in the reconstructed signals of the right panels of Figure 10. This demonstrates that these disturbances were propagated up to a higher initial altitude and can only be observed with the residual reconstructions.

[27] The method for reconstructing the $T_i$ residuals based on its main periodicities is a powerful tool in analyzing wave structure and propagation. In Figure 7, the periodicities seen are only projections in time onto one altitude. However, Figures 10 and 11 show the entire altitude range studied at the same time, making it possible to isolate specific periodicities in order to glean more detailed information about the dynamics of the ionosphere.

[28] In Figure 11, for periodicities between ~70 and 99 h, there appears to be a very slow tidal influence on these periodicities (see path that the dots follow). Finding the points where the temperature gradient switches signs produced this trace. These dots represent the inflection points of the temperature profiles. From Figure 11, the movement of the inflection points may be observed. The displacement of the temperature profiles could be responsible for observations of warming and cooling during an SSW as reported by Goncharenko and Zhang [2008], Funke et al. [2010], Kurihara et al. [2010], and Conde and Nicolls [2010]. The up and down movement of the inflection points is not likely a direct result of the SSW but could be a product of some other physical process, such as planetary wave-tide interactions. These interactions could provide a plausible explanation for the periodicities observed in Figure 8.

[29] In summary, Figures 10 and 11 indicate that the strongest periodicities found in midlatitude ion temperature are (in order of decreasing magnitudes) as follows: ~12 h, ~16–17 h, and ~30–40 h waves and ~10–13 d (>256 h) and 3–4 d (~70–98 h) components. Some of the similar periodicities are found in geomagnetic indices, in particular, ~16–20 h and ~30 h (Figure 9) and ~10–13 d. While we cannot completely rule out that some contributions to the observed variations in temperature are indirectly generated by geomagnetic forcing, our analysis approach is expected to remove dependency on geomagnetic activity, or at least to strongly decrease it. We also note that the spectral peaks in geomagnetic indices at ~16–20 h and ~30 h are relatively weak and mostly observed around 20 and 21 January, while periodic variations in $T_i$ are observed throughout the length of the campaign. In addition, $T_i$ disturbances due to magnetic activity are expected to remain within ~20–30 K for the observed level of geomagnetic activity (Figure 3), while observed periodicities in $T_i$ reach magnitudes of ±60 to 70 K. Finally, geomagnetic forcing is expected to influence ion temperatures throughout

Figure 8. Normalized power spectrum integration of the initial phase of the SSW event (period enclosed by the white lines of right panels of Figure 7). The white lines represent the strongest periodicities detected by altitude.

Figure 9. Power spectrum of the geomagnetic indices. The left panels show (from top to bottom) $ap$, $Dst$, and $AE$, while the right panels depict the periodicities observed throughout the campaign period (periods between ~4 and ~75 h).
Periodic variations in $T_i$ of these reasons, we focus on an interpretation of the observed ~250 km and strongly dissipate at higher altitudes. Due to all of these reasons, we focus on an interpretation of the observed $T_i$ variability as not related to the magnetic activity.

### 3.2. Disturbances with Tidal Periods

[30] Temperature variations with ~12 h periods are some of the strongest found in the data, reaching 60–90 K below ~250 km and strongly dissipate at higher altitudes. Due to all of these reasons, we focus on an interpretation of the observed $T_i$ variability as not related to the magnetic activity.

#### Figure 10.

Decomposition of the main periodicities that are less than 24 h. The left panel represents the $T_i$ residual oscillations at selected altitudes under study, while the isoline plots on the right shows the reconstructed periodicities. The white lines are $T_i = 0$ K, and the stratospheric temperature at 10 hPa for 90°N (light gray line) is superposed onto the isoline plots to help visualize the SSW event. The period ranges are 7.7–8.3 h, 9.9–12.9 h, and 14.1–18.2 h from the top row to the bottom row, respectively.

that the temporal development of a 12 h wave indicates high amplitudes in the 200–230 km region during the first part of the period (18–23 January), with rapid reduction by 24 January, and negligible 12 h amplitudes after 28 January. As solar flux gradually decreased during this period from $F10.7 = 82–85$ sfu on 18–24 January to $F10.7 = 73–75$ sfu on 29 and 30 January, tidal dissipation is expected to be weaker, enabling easier propagation to higher altitudes. Observations of stronger 12 h variations in the upper atmosphere during the higher solar flux conditions (18–23 January) indicate that the tidal amplitudes were significantly enhanced during this period (in comparison to 25–30 January) in order to counteract effects of higher solar flux. Thus, we conclude that the 12 h waves were strongly amplified in the midlatitude upper thermosphere during the first part of the campaign (18–23 January) and decreased to normal levels by the end of the campaign (28–30 January).

[31] We propose four possible mechanisms for the observed temporal variation in the 12 h wave. The first mechanism is related to the interaction of the planetary wave 1 and SW2 tide that is expected to enhance the nonmigrating
semidiurnal tide (SW1). As the planetary wave 1 amplitude is amplified prior to the peak of the SSW and strongly decreases during the mature stage of the SSW (see Figure 1d), the SW1 tide is expected to be stronger during the period of amplified PW and to become weaker as the planetary wave collapses. The second mechanism is related to variations in the zonal mean zonal wind (e.g., Figure 1c) to affect propagation of tides. Steining et al. [1997] have concluded that variations in the zonal mean wind typical for SSW conditions will enhance the lunar semidiurnal tide; similar enhancement can be expected for the solar semidiurnal tide. Another possible mechanism is related to the location of the radar’s field of view with regard to the disturbed polar vortex. In the study of the January 2009 SSW event, Yuan et al. [2012] reported significant anomalies in the midlatitude mesopause from the Fort Collins lidar (41°N) when the instrument’s field of view was within the disturbed stratosphere that extended from the polar region to the midlatitude region. Matthias et al. [2012] studied the mesospheric response to several SSW cases at high latitude (Andenes radar, 69°N, 16°E) and reported strong modulation of semidiurnal tides before and during SSW, which is consistent with earlier results [Hoffmann et al., 2007]. However, no clear relation in the temporal development of the 12 h tide relative to the onset of the SSW events was determined in their study. We consider the possibility that the temporal development of the 12 h tides during SSW will depend on the details of the disturbances in the polar vortex for a particular case and will strongly vary with longitude, as SSW events are inherently asymmetric. The fourth mechanism is related to variations in stratospheric ozone that undergoes significant changes during SSW at both low and middle-to-high latitudes [Randel, 1993; Pancheva et al., 2003; Sridharan et al., 2012; Goncharenko et al., 2012]. Ozone heating in the stratosphere due to solar UV flux is a major source of the semidiurnal tide, and short-term variation in the ozone density is a potential source of significant short-term variability in the semidiurnal tide. Future experimental and modeling studies are necessary to address the validity of the suggested mechanisms and their relative contributions to variability associated with SSW events.

Figure 11. The same as Figure 10 but for periodicities greater than 24 h. The period ranges are 23.8–44.1 h, 70.1–98.7 h, and periods greater than 256 h (from the top row to the bottom row, respectively). The dots in the middle panel represent a very slow tidal influence on the periodicities between 70.1 and 98.7 h.
(SW2) and amplification in the 8 h tide (TW3) during the peak of the warming [Fuller-Rowell et al., 2010, 2011a], with delayed increase in the SW2 after the SSW peak [Wang et al., 2011, 2012]. A study of the average tidal response to 23 SSW events using the Whole Atmosphere Community Climate Model (WACCM) found an enhancement in the SW2 tide and SW1 (nonmigrating) tide, as well as an enhancement in the semidiurnal lunar tide [Pedatella et al., 2012]. The Ground-to-topside model of Atmosphere and Ionosphere for Aeronomy (GAIA) simulation of a major 2009 SSW indicates amplification of the SW2 tide, with the largest increase reported in the low-latitude region, and a complex variation in the TW3 tide [Jin et al., 2012]. We note that the location of the Millstone Hill radar (42°N) is well suited for studies of these tidal variations, as models generally predict the largest enhancements in tidal modes during SSW in the middle- to high-latitudes. Observations reported in this study are generally consistent with amplifications of 12 h tides expected from the TIMEGCM, WACCM, and GAIA simulations but disagree with WAM results that predict a decrease in the 12 h waves.

[33] In addition to the strong 12 h periodicity, our results indicate a weaker wave with ~6.5–7.9 h periods (Figures 8 and 10). This periodicity attains a magnitude of ~20–50 K at ~200–230 km altitude and decreases to <10 K at higher altitudes. This wave could be a signature of a terdiurnal tide that is enhanced during the SSW and propagating from lower altitudes. As the wave propagates through the atmosphere with varying temperature and background wind, the wave period can change due to the Doppler shift and spectral broadening. Alternatively, the enhanced terdiurnal tide during SSW combined with the observed ~4 d wave (see Figure 8) would result in a 7.4 h secondary wave, which could explain the observed ~7 h periodicity.

[34] We note that contrary to simulations predicting an enhancement in the 24 h tide during SSW [Liu et al., 2010; Jin et al., 2012], our analysis does not show 24 h periodicities. It is possible that even if diurnal tides are enhanced, they dissipate at lower altitudes and do not directly propagate to the altitudes used in our analysis.

3.3. Disturbances with Nontidal and Multiday Periods

[35] It is well known that large-scale disturbances in the middle atmosphere can be described in terms of free Rossby waves [Salby, 1981; Madden, 2007]. Waves with periods of 5, 10, 16, and 25 d are frequently identified in the middle atmosphere, mesosphere, and lower thermosphere [e.g. Day and Mitchell, 2010; Sassi et al., 2012]. Vertical propagation of these waves is usually confined below the critical level near the mesopause [Sassi et al., 2002], but their signatures are often reported in the ionosphere as variations of ionospheric parameters with multiday periods [Forbes and Leveroni, 1992; Forbes, 1996; Aladell and Apostolov, 2003; Laštovička, 2006]. Our results do not show a 5 d periodicity, indicating that these waves did not directly propagate to the upper atmosphere, as expected. However, the nonlinear interactions between planetary waves and tides (primary waves) can generate two secondary waves whose frequencies are the sum and difference of frequencies of the primary waves [Teitelbaum and Vial, 1991], and some of our results could be interpreted in terms of these secondary waves.

[36] A strong variation with ~14–18 h period that reaches ~70–90 K below 270 km and decreases to ~20–30 K at higher altitudes (Figure 10) could be related to the quasi 2 d wave activity. The quasi 2 d wave (QTDW) is known to be enhanced in the summer mesosphere and lower thermosphere, with extension to the winter lower thermosphere [Wu et al., 1996]. Nonlinear interaction of the QTDW with tidal modes can introduce QTDW variation in the ionospheric electron density from equatorial [Chen, 1992] to middle latitudes [Forbes and Zhang, 1997] through the E region dynamo mechanism. The E region dynamo is not likely to affect ion temperature; thus, the QTDW oscillations can be seen in electron density but not in ion temperature. The 16 h periodicity shown in Figure 8 could be the product of an interaction between the QTDW and a semidiurnal tide (secondary waves are 9.6 h and 16 h). A portion of the ~16 h variation shown in Figures 7 and 8 can also be related to an ~17 h oscillation in geomagnetic indices shown in Figure 9, in particular, around 20 and 21 January.

[37] A significant enhancement (50–60 K) in variation with a ~1.5 d period is observed from 22 January to 26 January below 230 km (Figure 11). Pfister [1985] found that a baroclinic instability generated waves with periods between 1.4 and 3 d in the summer midlatitude mesosphere, and these waves are often interpreted as a part of QTDW oscillations. We note that the meteor radar collocated with the Millstone Hill ISR in Durham (43.1°N, 289.1°E) frequently observed QTDW variations in mesospheric winds, although summer season wave periods were closer to ~48 h, and wintertime mesospheric winds have peaks at 36–42 h [Clark, 1975]. The 1.5 d periodicity detected in our T, results is consistent with the 36–42 h periodicities in mesospheric winds reported by Clark [1975]. Pfister [1985] suggested that baroclinic instability may explain spectral peaks with these periods. However, Pfister [1985] emphasized processes in the summer mesosphere, and it is not clear if this interpretation is applicable to wintertime conditions.

[38] The weak 3–4 d wave (10–30 K) seen in Figures 8 and 11 could result from several different mechanisms. One of the normal atmosphere modes, a 5 d wave, can be observed in the mesosphere and lower thermosphere in a relatively broad spectrum of 4–6.5 d. Sassi et al. [2012] found this wave during some northern hemisphere winters, but not others. Talat et al. [2002] reported the 6.5 d wave in the tropical stratosphere from the UK Met Office data. In satellite data in the MLT region, this wave can be seen as a 5.5 d oscillation [Pancheva et al., 2010] or ~5 d oscillation [Wu et al., 1994]. Thus, it is possible that the ~4 d oscillations observed in our temperature data represent a Doppler-shifted ~5 d oscillation propagating from lower altitudes. Azeem et al. [2005] observed the 4 d wave in both the stratosphere and mesosphere, demonstrating an amplification in the 4 d wave preceding the SSW events in 1995 and 2002. Manney and Randel [1993] suggested that the generation of the 4 d wave is related to baroclinic and barotropic instabilities in the stratosphere. Thus, the weak 3–4 d oscillation in our data could also be generated by this 4 d planetary wave.

[39] Finally, a relatively strong (~30–40 K) oscillation with a period around ~250 h is observed at altitudes below ~250 km (Figure 11). The temporal development of this
periodicity indicates $T_i$ peaks on 23 and 25 January and $T_i$ minimums on 29 and 30 January. The aforementioned ~10 d periodicity in geomagnetic indices is expected to increase $T_i$ on 20 and 21 January and lead to a cooler $T_i$ on 25–27 January (Figure 1g). Thus, both the altitudinal and temporal developments of the 10 d periodicity in $T_i$ residuals lead us to believe that this periodicity is not related to the geomagnetic activity and could result from a 10–13 d wave originated in the lower atmosphere. The 10–13 d periodicity is one of the normal atmospheric modes that are frequently observed in ionospheric electron density [Altadill and Apostolov, 2003; Laštovička, 2006]. An $E$ region dynamo mechanism is usually invoked to explain transfer of the 10 d signature from the lower thermosphere to the ionosphere. Our results indicate that other mechanisms, either direct or indirect, are required to generate 10–13 d periodicities in temperature data.

[40] In addition to the 6.5 d wave, Talaat et al. [2002] also showed the presence of the 10 d wave and the 4 d wave in their data, suggesting that these waves were nonlinearly coupled with the 6.5 d wave or related through a common sourcing mechanism. The 4 and 10 d waves are present in the results shown in Figure 8; thus, if the 6.5 d is also present and it interacts with the semidiurnal tide, this would produce a 13 h secondary wave, which is observed at lower altitudes in our results. This mechanism could further enhance the 12–13 h oscillations expected from amplification of semidiurnal tides as discussed above.

[41] It remains an open question how planetary waves signatures can propagate from the stratosphere and into the upper thermosphere, as direct propagation of planetary waves is not expected from theory and modeling [Meyer and Forbes, 1997b; Pogoreltsev et al., 2007]. One possible mechanism that has some modeling and experimental support includes upward propagation of gravity waves and modulation of gravity waves by planetary waves [Meyer, 1999; Hoffmann et al., 2012], with subsequent propagation of PW-type variations to higher altitudes. Three-dimensional modeling studies with the extended GW parameterization of Yiğit et al. [2008] have demonstrated that small-scale GWs of lower atmospheric origin can propagate to $F$ region altitudes with appreciable dynamical and thermal effects [Yiğit and Medvedev, 2010, and references therein]. More recently, Yiğit and Medvedev [2012] showed that such effects vary significantly during SSW events due to changes in the zonal wind and GW dissipation.

[42] Another proposed explanation includes planetary wave-tide interactions that generate secondary waves with propagation characteristics different from the primary planetary waves. It has been suggested that the nonlinear interaction between planetary waves and tides causes nonlinear advection, transferring energy from the primary waves to the secondary waves [Forbes, 1996; Chang et al., 2011]. Further studies are required to determine whether these gravity waves and primary and secondary planetary waves can propagate to higher altitudes (up to 200–250 km). These scenarios explaining the periodicities observed in Figure 8 are just examples of possibly many scenarios involving interactions between planetary waves, tides, and gravity waves. These interactions seem to play an integral part in understanding the effects of SSW on the upper atmosphere and will require further analytic and observational studies to quantify.

[43] Several studies have suggested that excitation and propagation of planetary waves (free modes) are modified during SSW events, leading to amplification of these waves [Dowdy et al., 2004; Pancheva et al., 2009]. Matthias et al. [2012] have demonstrated increase in the planetary wave activity in the mesosphere from both ground-based and satellite observations during several SSW events, including the period in this study. Recently, Sassi et al. [2012] used the Navy Operational Global Atmospheric Prediction System-Advanced Level Physics and High Altitude global meteorological analyses in the 0–92 km altitude range to investigate the behavior of Rossby normal modes (4, 5, 10, 16, and 25 d) during several winters. Although significant wave activity was found, the amplitudes of the modes were not enhanced during sudden stratospheric warmings (disturbed winter) in comparison to the nonwarmings (quiet winter). Our analyses suggest that signatures of normal modes can be propagated to the upper atmosphere, at least for the period under study. It is not known at this point whether these signatures are strongly affected by the SSW phenomenon or whether similar variability is a regular feature of the upper atmosphere that can be observed during other times.

4. Conclusions

[44] We presented observations of the periodic variations in upper atmospheric ion temperature between approximately 200 and 400 km that were obtained by the Millstone Hill ISR (42.6°N, 288.5°E) during the experimental campaign of 18–31 January 2010. The campaign took place during the mature stage of sudden stratospheric warming, providing the opportunity to investigate the upper thermospheric response to a strong disturbance in the middle atmosphere. To delineate ionospheric variations forced from below from those forced by solar and geomagnetic activity, we constructed an empirical model of ionospheric temperature using wintertime data collected in 2007–2009. We analyzed the residuals between the observed temperature and the model’s estimates to investigate periodicities on time scales from several hours to 10–13 d. The primary results of this study are as follows:

1. The ion temperature in the upper atmosphere during the winter season of 2007–2009 is significantly lower than during earlier periods of low solar activity. This is consistent with multiple reports of a colder thermosphere during the recent solar minimum. The largest differences are observed during the daytime and reach ~100 K.
2. A strong temperature variation with ~10–13 h period and ~50–90 K amplitude is found in the altitude range of 200–250 km and interpreted as an evidence of an enhanced semidiurnal tide that propagated to the upper thermosphere. This enhancement is observed during the first part of the campaign (18–24 January) and is likely to be associated with the sudden stratospheric warming.
3. A weaker 20–50 K variation with a period of ~6.5–8 h is retrieved from the data below ~250 km altitude. This
variation could result from the enhancement of the terdiurnal tide during sudden stratospheric warming.

4. Ion temperature variations of different strength are found at multiple nontidal periods, with the largest variations reported for periods of 16–17h, 30–40h, 10–13d, and 3–4d. This variability appears to be driven by planetary waves with periods of 4d, 5d, and 10–13d.

5. The oscillations with a 3–4d period and a 10–13d period can result from upward propagation of planetary waves with these periods. The oscillation with 16–17h period could result from the nonlinear interaction of the quasi 2d wave with a semiannual tide.

6. Analysis of data for one experimental campaign limits our ability to distinguish whether temperature oscillations with nontidal periods (16h, 30–40h, 3–4d, and 10–13d) are solely associated with sudden stratospheric warming or are part of the regular ionospheric variability.

7. Multiple studies of ionospheric disturbances with planetary wave periods are focused on variations in electron density and interpret their findings using the $E$ region dynamo mechanism, as planetary waves are not expected to propagate to $F$ region altitudes. However, our observations of ion temperature variations with planetary wave periods (and periods resulting from interaction of planetary waves with tides) do not necessarily require a distinct $E$ region dynamo driver.

These results indicate that such waves can propagate up to ~250 km, at least in the case of deep solar minimum, and dissipate at altitudes above ~250 km. To the best of our knowledge, this is the first experimental evidence of direct (or indirect) propagation of planetary waves to the upper thermosphere.

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