Observations of Quasi-Two-Day wave by TIMED/SABER and TIMED/TIDI

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[1] Seasonal and interannual variations of the Quasi-Two-Day wave \( s = -3 \) (W3) and \( s = -4 \) (W4) modes were studied with global temperature and wind data sets during 2002–2012, observed respectively by the Sounding of the Atmosphere using Broadband Emission Radiometry (SABER) and TIMED Doppler Imager (TIDI) instruments onboard the Thermosphere Ionosphere and Mesosphere Electric Dynamics (TIMED) satellite. The amplitudes of W3 and W4 are significantly enhanced during austral and boreal summer respectively. Strong W3 amplitudes are observed during January 2006 in all three components of temperature, meridional wind, and zonal wind. This is most likely related to the intensive winter planetary wave activity that led to a strong sudden stratosphere warming (SSW) event. The maximum amplitudes of W4 during the 10 years are \( \sim 8–9 \) K, \( \sim 40 \) m/s, and \( \sim 20 \) m/s for temperature, meridional and zonal components respectively, nearly half as large as those of W3, with \( \sim 15 \) K, \( \sim 65 \) m/s, and \( \sim 35 \) m/s. In January 2008 and 2009, unusually weak W3 but strong W4 oscillations were observed, corresponding to the much weaker summer easterly jets (westward wind) than those in other years. This suggests that relatively weak summer easterly may not be able to provide sufficiently strong barotropic/baroclinic instability to amplify W3 but is favorable for the amplification of W4. The weaker magnitude values, lower peak heights, and longer life intervals of W4 than those of W3 suggest that the W4 may suffer a greater damping rate than the W3. The observations of W4 show good agreement with Rossby-gravity \((4, 0)\) mode, which is more easily trapped in both latitude and altitude because of its lower group velocity than that of Rossby-gravity \((3, 0)\) mode.


1. Introduction

[2] As a significant oscillation in the mesosphere and lower thermosphere (MLT) region, the Quasi-Two-Day wave (QTDW) has been studied with temperature and wind data sets observed by both space-based and ground-based instruments [Rodgers and Prata, 1981; Plumb, 1983; Burks and Leovy, 1986; Wu et al., 1993; Herman et al., 1999; Thayaparan et al., 1999; Lima et al., 2004; Pancheva et al., 2004; Li et al., 2008; Tunbridge et al., 2011]. The QTDWs are mostly observed to be westward propagating with zonal wave number 3 (W3) or 4 (W4), and the W3 is generally much stronger (nearly twice) than W4.

[3] Two primary theories, i.e., Rossby-gravity \((3, 0)\) mode [Salby, 1981a, 1981b] and atmosphere barotropic/baroclinic instability [Plumb, 1983; Salby and Callaghan, 2001], have been proposed to interpret the origin of the W3. The latitudinal and vertical distributions of W3 are governed by the Rossby-gravity normal mode [Salby, 1981a], while the amplitude and duration of W3 could be significantly modified by atmosphere instability [Palo et al., 2007]. The W3 could also be a subharmonic resonance excited by atmospheric solar tides [Walterscheid and Vincent, 1996; Hecht et al., 2010; McCormack et al., 2010], and the nonlinear interactions between the W3 and migrating diurnal solar tides have been proven to be significant in the amplification of the W3 in January 2006 [McCormack et al., 2010]. The W3 are primarily observed during January in the Southern Hemisphere [Palo et al., 2007; Yue et al., 2012].

[4] Rodgers and Prata [1981] first reported a westward travelling planetary wave with zonal wave number 4 and a period of \( \sim 1.7 \) days when analyzing radiance data observed...
by the Pressure Modulator Radiometer instrument onboard the Nimbus 6 satellite, which was also confirmed by Plumb [1983] with a one-dimensional model under summer mesospheric flow conditions. This new type of QTDW is believed to be the result of summer easterly jet baroclinic instability [Plumb, 1983; Burks and Leovy, 1986]. Salby and Callaghan [2001] proposed an alternate interpretation for the W4, which is that the observed W4 is consistent with Rossby-gravity (4, 0) normal mode. The (4, 0) mode grows much slower than the (3, 0) mode under a strong summer easterly representative of January conditions. However, they are amplified equally fast under a weak summer easterly representative of July conditions. In other words, zonal wind in July is more favorable for the amplification of W4 than that in January.

[5] Significant seasonal and interannual variations of both W3 and W4 have been studied with satellite temperature observations [Wu et al., 1996; Garcia et al., 2005; Palo et al., 2007; Tunbridge et al., 2011]. It is found that the QTDWs exhibit remarkable interannual variability but with stable vertical and global distributions. Satellite wind observations of the W3 have been presented earlier by Wu et al. [1993] and Ward et al. [1996] with UARS/HRDI and UARS/WINDII horizontal wind measurements during 1992–1993 and by Limpasuvan and Wu [2009] and Limpasuvan et al. [2005] with Aura Microwave Limb Sounder (MLS) line of sight wind observations. The analysis of UARS/HRDI horizontal wind in January 1992/1993 suggests that the W3 maximizes between 90 and 100 km with amplitudes of ~60 and ~30 m/s in meridional and zonal winds, respectively [Wu et al., 1993]. Until now, there was no report on the global distributions of the W4 in horizontal winds and its seasonal and interannual variabilities.

[6] In this paper, we focus on the long-term variations of the W3 and W4 using both global temperature and horizontal wind data sets. The Sounding of the Atmosphere using Broadband Emission Radiometry (SABER) and TIMED Doppler Imager (TIDI) instruments onboard the Thermosphere Ionosphere and Mesosphere Electric Dynamics (TIMED) satellite have measured the global temperature and the zonal and meridional winds in the MLT region for nearly 11 years, providing us an opportunity to examine the seasonal and interannual variations of the W3 and W4 in temperature and horizontal wind. The SABER and TIDI data sets and the data analysis method are briefly described in section 2. In section 3, we first discuss the aliasing issues when analyzing TIMED satellite observations. Then the seasonal and interannual variations of both W3 and W4 are presented and discussed. A summary follows in section 4.

2. Data Sets and Data Analysis

[7] The SABER and TIDI instruments onboard the TIMED satellite have been providing temperature and wind observations in the MLT region, respectively, since late January 2002. A number of SABER temperature validation studies have been conducted and reported in the literature, including Mertens et al. [2004], Rensberg et al. [2008], and Garcia et al. [2008]. Also, Stevens et al. [2012] gave insight regarding SABER high-altitude temperature uncertainties through comparisons with measurements made with the SOFIE instrument on the AIM satellite. Overall, the absolute temperature accuracies range from ±0.5 K below 75 km, ±1–2 K from 75 to 95 km, to ±4 K at 100 km. Both ascending and descending data sets of Level2A version 1.07 are utilized here to retrieve W3 and W4 below 95 km from January 2002 to January 2012. The missing data gaps in our analysis are mainly caused by cycling the power to the SABER cooler, which has to be done periodically to remove ice buildup on the detectors or due to the “Safehold” mode of TIMED spacecraft when there are spacecraft issues.

[8] A new zero-wind implementation is used in retrieving the NCAR-processed version 0307A of P9 line TIDI horizontal wind with a vertical resolution of ~2 km [Killeen et al., 2006; Wu et al., 2008a]. This version of TIDI horizontal wind data set is successfully utilized in studying migrating and nonmigrating tides [Wu et al., 2008a, 2008b] and semidiurnal tides [Wu et al., 2011]. The TIDI instrument was designed to measure horizontal vector wind with accuracies of ~3 m/s [Killeen et al., 2006]; however, the systematic errors are difficult to be determined [Wu et al., 2008a]. Oberheide et al. [2006] took the error numbers to be 30 m/s for an individual data point during the day and double that during the night. The University of Michigan has been reporting gradual improvement of the TIDI data over the past years. Both ascending and descending TIDI cold-side data sets between 85 and 105 km during 2002–2012 are analyzed. The performance of TIDI instrument is nominal, except when the background white light is higher than the expected level or ice deposits on optical surfaces [Killeen et al., 2006].

[9] At a given latitude and altitude, the travelling planetary waves may be expressed as a sum of cosine and sine terms with different wave numbers and frequencies and the background mean field C:

\[ y = A \cos[2\pi(\sigma \times t + s \times \lambda)] + B \sin[2\pi(\sigma \times t + s \times \lambda)] + C \]  (1)

where \( \sigma \) and \( s \) are the frequency and zonal wave number of a certain planetary wave, respectively, \( t \) and \( \lambda \) are the UT time and longitudes of the satellite samplings, respectively. Then a least squares method is used to fit the magnitudes and uncertainties of \( A, B, \) and \( C \). The amplitude \( R(\sigma, s) \) of wave \( (\sigma, s) \) could be expressed as:

\[ R^2(\sigma, s) = A^2(\sigma, s) + B^2(\sigma, s) \]  (2)

[10] To calculate the spectra, we increase the frequencies from 0.01 to 1.0 cpd in steps of 0.01 cpd and the wave numbers from −6 to 6 in steps of 1. To extract the amplitudes of the QTDW, we fix the wave number at −3 and −4 for the W3 and W4, respectively, and only increase the frequency \( \sigma \) from 0.35 cpd (2.8 days) to 0.65 cpd (1.5 days) with a resolution of 0.01 cpd. The strongest amplitude and the corresponding frequency are taken as the magnitude and frequency of the QTDW in every fit, similar to the manner applied by Tunbridge et al. [2011]. In our analysis, each calculation is performed on a time interval of 8 days stepped forward by 2 days for SABER temperature observations, while a 15 day temporal window stepped by 3 days is used for TIDI wind analysis. The SABER temperature data sets are also averaged over 4 km stepped by 2 km in altitude and
then analyzed within a band of 10° stepped by 5° in latitude. The TIDI wind observations are analyzed within a band of 20° stepped by 5° in latitude. The TIDI instrument has much more sparse sampling points than the SABER instrument; therefore, much broader temporal and latitudinal windows are used to reduce the statistical uncertainties in horizontal wind. The centers of these latitude and altitude intervals are taken as the spatial reference in every fitting. We visually examine the least squares fitting by overplotting the raw data with the reconstructed time series from least squares fitting, and a reasonably good agreement is found.

The UK Met Office (UKMO) assimilation data are also used to provide monthly mean of zonal mean zonal wind to investigate the strength of summer mesospheric easterly jets. The UKMO data assimilation system is used for operational weather forecasting, which provides daily zonal wind data from the ground to 0.1 hPa (or 0.3 hPa during certain years). The background information of the UKMO assimilation system is discussed in detail by Swinbank and O’Neill [1994] and Swinbank and Ortland [2003].

3. Results and Discussion

3.1. Aliasing and Uncertainties

[12] We first consider the aliasing issues of the analysis method. Synthetic global waves with constant amplitude and a 2 day period but different wave numbers are reconstructed at the SABER and TIDI sampling points, and then the new data sets are analyzed to examine the aliasing to other wave numbers and frequencies. This is similar to the method used by Tunbridge et al. [2011] for analyzing QTDW aliasing with Aura MLS temperature observations.

Figure 1 shows the aliasing spectrum of W3 and the percentage of W3 aliased from the westward QTDW $s = -2$ mode (W2), W4, and the eastward QTDW $s = +2$ mode (E2) obtained with SABER temperature and TIDI wind observations.

Figure 2. (a) Aliasing signals from W3. SABER sampling points during 22–29 January 2005 between 30°S and 40°S are utilized in the analysis. (b, c) Amplitudes of W3 aliased from W2, W4, and E2 are obtained with SABER temperature sampling points for Figure 1b and TIDI wind sampling points for Figure 1c. The red (blue) plots are obtained with SABER or TIDI observations when the sampling points cover the Southern (Northern) Hemisphere. The sampling patterns of SABER and TIDI instruments are different in separate satellite yaw maneuver period.

Figure 2. (a) Aliasing signals from W4. SABER sampling points during 1–8 January 2005 between 25°N and 35°N are utilized in the analysis. (b, c) Amplitudes of W4 aliased from W2, W4, and E3 are obtained with SABER temperature sampling points for Figure 2b and TIDI wind sampling points for Figure 2c. The red and blue plots are the same as those in Figure 1.
horizontal wind sampling points. It is clear from Figure 1a that only a large amount of E2 could be aliased from W3 (and vice versa). The aliasing issues between W3 and E2 due to the analysis of satellite observations have also been discussed earlier by Tunbridge et al. [2011] and Wu et al. [1995]. Figures 1b and 1c show the amplitude percentages of W2, W4, and E2 aliased to W3. The aliased signals of W3 from W2 and W4 are trivial compared with that from E2. The percentages of aliased amplitudes in temperature from E2 to W3 increase from ~30% at 50° in one hemisphere to ~90% at 50° in the other hemisphere. Meanwhile, the percentages of aliasing from E2 to W3 in horizontal wind increase from ~10% at 50° in one hemisphere to ~80% at 50° in the other hemisphere.

[14] Figure 2 shows the aliasing spectrum of W4 and the percentages of W4 aliased from W2, W3, and eastward QTDW \( s = +3 \) mode (E3). It is clear from Figure 2a that a large amount of E3 could be aliased from W4 (and vice versa). Tunbridge et al. [2011] had also showed that the W4 and E3 could be aliased to each other with Aura MLS satellite observations. Figures 2b and 2c show the percentages of W2, W3, and E3 aliased to W4. The aliased signals of W4 from W2 and W3 are negligible compared with that from E3. The percentage of E3 in temperature aliased to W4 increases from ~30% at 50° in one hemisphere to ~90% at 50° in the other hemisphere, while the aliasing percentage of W4 from E3 in horizontal wind increases from ~10% at 50° in one hemisphere to ~80% at 50° in the other hemisphere. Since there is no real mesospheric E3 signal reported yet, the aliasing effect of E3 on W4 could be neglected in our analysis.

[15] We also consider the one-sigma statistical uncertainties of W3 and W4 in our analysis, similar to the work done by Wu et al. [2008a, 2008b] in studying diurnal tides and by Tunbridge et al. [2011] in studying QTDW. Figure 3a shows the uncertainties of W3 at 84 km with an 8 day temporal and 10° latitudinal window. The temperature uncertainties are clearly less than 1 K most of the time, while the uncertainties at 95 km are normally ~1–1.5 K all the time. The temperature amplitude uncertainties of W4 are comparable to those of W3. Further, since the QTDW in temperature is fairly weak below 50 km, we only plot the analysis results between 50 and 95 km hereafter. The wind amplitude uncertainties for W3 and W4 are mostly less than 5 m/s with a 15 day temporal and 20° latitudinal window, as shown in Figures 3b and 3c for W3 at 92.5 km. Generally, the maximum amplitude uncertainties are also ~5 m/s at other altitudes.

3.2. The Westward \( s = -3 \) Mode

[16] Figure 4 shows the frequency-wave number spectra of W3 in January 2006 obtained with SABER temperature observations between 82 and 86 km and between 30°S and 40°S during 14–21 January, (b) TIDI meridional wind observations at 92.5 km between 15°S and 35°S during 10–24 January, and (c) TIDI zonal wind observations at 92.5 km between 40°S and 60°S during 10–24 January. W3 mode peaks at \( f = -0.55 \) cpd in all three components. E2 also stands out of the spectrum with frequency at ~0.45 cpd.
Figure 5. Latitudinal and vertical structures of W3 in (a) SABER temperature, (b) TIDI meridional wind, and (c) TIDI zonal wind. Observations between the same episodes as those of Figure 4 are used in the analysis. The contour intervals are 1 K and 5 m/s for temperature and horizontal wind, respectively.

Figure 6. Temporal and vertical structures of W3 observed by SABER temperature between 30°S and 40°S during 2003–2012. The contour interval is 1 K. The blanks in late January 2003 and early January 2008 represent missing data.

40°S during days 14–21, TIDI meridional wind observations at 92.5 km between 15°S and 35°S during days 10–24, and TIDI zonal wind observations at 92.5 km between 40°S and 60°S during days 10–24. W3 and E2 dominate the whole spectrum at frequencies $f \approx 0.55$ and $\approx 0.45$ cpd, respectively, in all the three components. Wave number-frequency spectra in
January during other years (not shown) exhibit similar QTDW features as displayed in Figure 4 but with different magnitudes and frequencies. As shown in Figure 1b, 70% of the amplitude of E2 could alias to W3 at 35°S for SABER temperature (and vice versa). Similarly, 30% and 70% of the amplitude of E2 could alias to W3 at 25°S and 50°S for TIDI meridional and zonal winds, respectively (and vice versa), as shown in Figure 1c. The amplitudes of E2 in temperature and horizontal wind in our analysis are comparable to the energy aliased from W3. Moreover, the E2 mode is generally considered as a polar mesospheric phenomenon of the winter hemisphere according to the previous observations [Prata, 1984; Nozawa et al., 2003; Sandford et al., 2008; Tunbridge et al., 2011]. In this case, the E2 in our analysis is difficult to distinguish from an aliased signal of W3 due to the analysis method and real signal, although Palo et al. [2007] and Pedatella et al. [2012] declared that this E2 is due to nonlinear interaction between W3 and the migrating diurnal tide. The W3 at middle- and low-latitude regions of summer hemispheres is most likely a real wave.

[17] Shown in Figure 5 are the global and vertical structures of W3, obtained with SABER temperature during January 14–21 and TIDI horizontal winds during January 10–24. W3 in temperature maximizes at ~84 km and ~40°S with amplitude of ~15 K and a second peak of ~10 K at ~68 km and ~35°S. The perturbations reach minimum at the equator, with only weak amplitudes above 80 km at the middle-latitude regions of the winter hemisphere. Figure 5b shows that W3 in meridional wind reaches maximum at 92.5 km and ~20°S, with amplitude of ~65 m/s. Different from the global structure of temperature, the W3 in meridional wind propagates across the equator to as far as 40°N in the winter hemisphere. Though the W3 in zonal wind is not completely exhibited in Figure 5c because of the limited satellite observations, we could still see that the zonal wind amplitudes are nearly half as strong as those of meridional wind. It is clear that W3 in zonal wind maximizes at higher-latitude regions of the summer hemisphere than meridional wind, with maximum amplitude of ~35 m/s near 50°S. The wave perturbations reach minimum at the equator,

Figure 7. The same as Figure 6 but for temporal and global variations of W3 obtained with SABER temperature observations between 82 and 86 km.
with weak amplitudes in the winter hemisphere, which are similar to that in temperature. The W3 in both meridional and zonal winds reach maxima at ~92.5 km. In general, both the amplitudes and the spatial structures of W3 in our analysis obtained with SABER temperature and TIDI horizontal wind observations are comparable with the results as reported in the previous literatures [Wu et al., 1993; Pancheva et al., 2006; Palo et al., 2007; McCormack et al., 2009; Hecht et al., 2010].

[18] The W3 in both temperature and horizontal wind are amplified mainly in January and early February, and this major W3 event has been extensively studied [Wu et al., 1993; Limpasuvan et al., 1995; Ward et al., 1996; Wu et al., 1996; Salby and Callaghan, 2001, 2003; Limpasuvan et al., 2005; Palo et al., 2007; McCormack et al., 2009; Tunbridge et al., 2011; Yue et al., 2012]. In the Southern Hemisphere, no significant signals of W3 occur at other times of the year except during the primary January event. In the Northern Hemisphere, weak temperature and meridional wind wave amplitudes are also observed in late July and early August with maxima located at middle- and low-latitude regions. However, the July/August amplitudes for temperature and meridional wind reach only 4 K and 20 m/s, respectively, much smaller than the January/February amplitudes. The W3 during July/August are also observed by Wu et al. [1996] and Baumgaertner et al. [2008], which are also much weaker than the January/February event. Therefore, we will focus on the period of January/February to study the interannual variations of W3 during 2003–2012.


Figure 8. Temporal and vertical variations of W3 obtained with TIDI meridional wind observations between 15°S and 35°S during 2003–2012. The contour interval is 5 m/s. The blanks in early January 2008 and January 2010 represent missing data.
between 15°S and 35°S and at 92.5 km, respectively. The W3 in middle January 2006 is much stronger than that in any other year, with amplitude of ~65 m/s at ~25°S and ~92.5 km, followed by the amplitudes of 45–50 m/s in January 2003, 2004, and 2007. In 2005 and 2011, the amplitudes of W3 in meridional wind reach ~30 and ~35 m/s, and the minimum amplitudes in 2008 and 2009 are only ~20 m/s.

The temporal and vertical variations of W3 in zonal wind during January 2003–2012 are shown in Figure 10, obtained with TIDI observations between 40°S and 60°S. The strongest W3 in zonal wind occurred in 2006 and 2011, with amplitudes of ~35 m/s at 92.5 km, followed by a value of 30 m/s in 2003. The amplitudes of W3 in zonal wind never exceed 25 m/s in other years.

One of the most notable features of the interannual variations of W3 is that it reaches minimum in January 2008 and 2009 with amplitudes of ~5 K, ~20 m/s, and less than 10 m/s for temperature, meridional wind, and zonal wind, respectively. The UKMO monthly mean of zonal mean zonal wind in January suggests a weaker summer easterly jet during these two years compared with that during other years, as shown in Figure 11. Since the summer easterly jet provides a favorable condition for the reversal of the latitudinal gradients of the potential vorticity, which is a prerequisite for the occurrence of atmospheric barotropic/baroclinic instability [Liu et al., 2004]. The weak summer easterly wind shear in 2008 and 2009 may not be strong enough to form sufficient baroclinic/barotropic instability [Plumb, 1983] for the amplification of the W3 [Burks and Leovy, 1986; Wu et al., 1996]. McCormack et al. [2010] performed assimilation studies of the amplification of W3 through nonlinear interaction between the migrating diurnal tide via $(1, \cos^6)$ diurnal tide in 2006 and 2008. They found that the summer easterly jet, W3, and diurnal tide $(1, -6)$ were weaker in January 2008 compared with those in January 2006. Walterscheid and Vincent [1996] found that nonlinear excitation of $(1, -6)$ mode is only favored when the background zonal wind reaches or exceeds 58 m/s. It is possible to observe a weak (strong) W3 in January 2008 (2006) with a relatively weaker (stronger) summer easterly jet according to the tidal generation theory. However, the periods of W3 we derived from SABER temperatures and

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**Figure 9.** The same as Figure 8 but for temporal and global variations of W3 obtained with TIDI meridional wind at 92.5 km.
TIDI winds do not always agree with the 48 h phase-locking features recommended by the tidal generation mechanism, as are shown in Figure 4. In fact, the analysis of 11 year upper mesospheric radar wind data at 35°C by Harris [1994] also suggested that the QTDW is not always precisely phase locked with solar local time. Hecht et al. [2010] declared that the phase-locked QTDWs only refer to those waves whose mean periods are approximating 48 h.

It is also very interesting to see the coincidence of the minima of W3 with the solar minima during 2008–2009. With LF range D1 radio wind observations at the middle latitude of the Northern Hemisphere, Jacobi et al. [1997] found that the amplitudes of QTDW were larger during solar maximum years, likely due to the stronger summer easterly in lower mesopause region, which favors the amplification of QTDW. Keuer et al. [2007] found that the summer easterly was stronger during the solar maximum periods of 1990–1992 and 2000–2002 than during the solar minimum years of 1995–1997 with MF-radar data set at (54.6°N, 13.4°E). However, a continuous 12 year study of the upper mesosphere wind at (35°C, 118°E) during 1980–1991 showed that the QTDW in meridional wind was even stronger in the solar minimum years of 1985–1987 than in the neighboring years [Harris, 1994]. Currently, we are not sure whether this is an evidence of a true geophysical connection between the amplitudes of W3 and solar cycle or merely a coincidence. Possible modulation of the W3 by the 11 year solar cycle needs further analysis with a longer data set and model simulation. Longer observations of the austral summer easterly are also needed to exhibit the relationship between the strength of the austral summer easterly and the solar activities.

The W3 in January 2006 is the only event when the temperature, meridional wind, and zonal wind oscillations are all very strong, with amplitudes of ~15 K, ~65 m/s, and ~35 m/s, respectively. These enhanced perturbations coincide with a strong SSW event at high-latitude regions of the Northern Hemisphere during this period. McCormack et al. [2009] studied the evolution of the W3 in January 2006 with a forecast assimilation model. High-level winter

Figure 10. The same as Figure 8 but for TIDI zonal wind with observations between 40°C and 60°C. The blanks in January 2010 represent missing data. The amplitudes of January 2008 and 2009 are below 10 m/s which are not plotted.
Figure 11. UKMO monthly zonal mean zonal wind in January during 2003–2012. The contour interval is 10 m/s.

Figure 12. Frequency-wave number spectra of W4 obtained with (a) SABER temperature observations between 70 and 74 km and between 30°N and 40°N on days 209–216, 2006, (b) TIDI meridional wind observations at 90 km between 25°N and 45°N on days 212–226, and (c) TIDI zonal wind observations at 90 km between 25°N and 45°N on days 212–226. The W4 mode peaks at $f=-0.52$ cpd in all the three components, accompanied by the E3 mode.
planetary waves were observed before the occurrence of the SSW, which were believed to be significant in triggering and amplifying the QTDW [Wu et al., 1996; Limpasuvan et al., 2000; Salby and Callaghan, 2001]. Salby and Callaghan [2001] found that inertial instability took place in the equatorial region where winter planetary waves were absorbed when propagating into the summer hemisphere, which could reinforce the mean meridional circulation and preserve a favorable atmosphere condition for the amplification of W3. In January 2006, the enhanced northward component of the residual meridional circulation could further affect the background zonal wind and the forcing of W3 [McCormack et al., 2009]. The horizontal momentum advection induced by planetary wave breaking in the
subtropical upper stratosphere and lower mesosphere favors the amplification of the W3 through barotropic instability. As such, the enhanced W3 oscillation during January 2006 is likely related to strong planetary wave activity in the Northern Hemisphere. We should also note that the SSW also occurred in minimum W3 years of January 2008 and 2009, with the latter being the strongest and most prolonged on record [Manney et al., 2009]. However, the summer easterly jet was relatively weaker in January 2008 than in January 2006 [McCormack et al., 2010]. We speculate that the amplification of the W3 will most probably be enhanced during the SSW event only if the background zonal wind favors the propagation of W3 (e.g., strong summer easterly jet). The influence on the amplification of W3 will be limited during the years with unfavorable background conditions for W3 (e.g., weak summer easterly jet). Model simulations are needed to study the joint influence of the SSW and summer easterly jet on the amplification of W3.

3.3. The Westward $s = -4$ Mode

[23] Least squares fitting spectra of SABER temperature, TIDI meridional wind, and TIDI zonal wind observations in August 2006 are shown in Figure 12. The temperature spectrum is obtained with SABER observations between 70 and 74 km and between 30°N and 40°N on days 209–216. The wind spectra are obtained with TIDI observations at 90 km on days 212–226 between 25°N and 45°N for both meridional and zonal wind components. The W4 mode peaks at $f = -0.52$ cpd, accompanied by a secondary E3 mode. Although nonlinear interaction between the migrating diurnal tide and the W4 could be applied to interpret the origin of the E3 mode [Teitelbaum and Vial, 1991], a large amount of the power of E3 could also be aliased from W4 at a middle-latitude region as shown in Figure 2. Since there is no real mesospheric E3 signal reported yet, the aliasing effect of E3 on W4 could be neglected in our analysis.

[24] Figure 13a shows the global and vertical structures of W4 obtained with SABER temperature on days 209–216 and TIDI meridional wind on days 212–226 in 2006. The W4 in temperature is mainly limited to the summer (northern) hemisphere with maximum amplitude of ~8 K at ~70 km and ~35°N, and the perturbations are negligible below 60 km or above 95 km. Figure 13b shows that the W4 in meridional wind reaches maximum at 90 km and ~35°N with amplitude of ~40 m/s. Similar to the global structure of W4 in temperature, the W4 in meridional wind does not extend across the equator to the winter hemisphere. The W4 in zonal wind

Figure 15. The same as Figure 14 but for temporal and global variations of W4 obtained with SABER temperature observations between 70 and 74 km.
(not shown) is much weaker than that in meridional wind in the middle-latitude region of the summer hemisphere.

[25] The W4 mode is observed to be amplified mainly in July/August in the Northern Hemisphere in our analysis and in the previous literatures [Tunbridge et al., 2011; Wu et al., 1996; Salby and Callaghan, 2001]. Figures 14 and 15 show the temporal variations of W4 in temperature during July/August 2002–2011 with observations between 25°N and 35°N and between 70 and 74 km, respectively. The temperature perturbations of W4 are completely limited within the summer hemisphere and are usually amplified twice during July/August. The most robust W4 in temperature occur during days 200–220 (late July and early August) in 2006 and during days 180–200 (early and middle July) in 2009 with amplitudes of 7–9 K. The amplitudes of W4 reach 5–6 K in 2002, 2008, and 2010 and are less than 4 K during other years. Temporal variations of the W4 in TIDI meridional wind during July/August 2002–2011 are shown in Figures 16 and 17 with observations between 30°N and 50°N and at 90 km, respectively. The W4 in August 2006 and 2004 reach ~30–40 m/s, which are stronger than those in other years with amplitudes of 20–30 m/s (or even less in August 2003). The amplitudes in zonal wind are less than ~20 m/s, even during the years with strong W4 event.

[26] Generally, the W3 dominates the QTDW spectrum in January in the Southern Hemisphere, and the W4 dominates the QTDW spectrum in July and August in the Northern Hemisphere. However, strong W4 signals dominate the QTDW spectrum of the Southern Hemisphere in January 2008 and 2009, as shown in Figure 18a. Shown in Figures 18b and 18c are the global and vertical distributions of this strong W4 event during January 2009 in temperature and meridional wind, respectively. The amplitudes of W4 reach maxima of ~7 K at 74 km and ~25 m/s at 92.5 km for temperature and meridional wind, respectively. This austral summer event is as strong as that in boreal summer as shown in Figures 14–17. Similarly, both the temperature and meridional wind QTDWs are limited mainly to the summer hemisphere. A secondary maximum of ~5 K at 58 km in temperature is also apparent in Figure 18b. Note that the austral summer easterly jets in 2008 and 2009 are weaker than those in other years (Figure 11). This suggests that the W4 is more easily amplified with a relatively weak summer easterly condition, likely due to the lower critical level, which is essential to the amplification of W4 through wave-flow interactions [Offermann et al., 2011].
The W4 shows many different features from the W3. First, the W4 begins to amplify from late June and fades away before the middle of August in the Northern Hemisphere, while the W3 is observed to be strongest in January in the Southern Hemisphere. The UKMO zonal mean zonal wind also shows that the boreal summer easterly is weaker than that in austral summer (not shown). In other words, the weak July and August summer easterly is favorable for the amplification of W4, while the strong January summer easterly is favorable for the enhancement of W3. This is further confirmed by the extraordinary strong W4 wave amplitudes during austral summer in 2008 and 2009.

Figure 17. The same as Figure 16 but for temporal and global variations of W4 obtained with TIDI meridional wind at 90 km.

Figure 18. (a) SABER temperature spectrum and (b, c) global and vertical distributions of W4 in temperature for Figure 18b and meridional wind for Figure 18c in January 2009. SABER temperature observations during 12–19 January and TIDI meridional wind during 9–23 January were used in the analysis. The contour intervals are 1 K and 2 m/s for temperature and meridional wind, respectively.
2009 as discussed above. Second, the W4 in temperature and wind maximize, respectively, at ~70–75 and ~90 km, which are lower than the peaks of W3 at ~80–85 and ~92.5 km. Third, the amplification of W4 could last a longer time interval of nearly 2 months than that of W3 with less than 1 month, and the W4 is amplified more than once during its lifetime. Fourth, the amplitudes of W4 in both temperature and horizontal wind are weaker than that of W3. The W4 is mostly confined in the summer hemisphere, while the W3 could propagate from the summer hemisphere to 30°N in the winter hemisphere during January. The smaller magnitude values, lower peak heights, and longer life intervals of the W4 than those of the W3 may suggest that the W4 suffers a greater damping rate than the W3.

[32] The simulation results by Salby and Callaghan [2001] showed that a weak summer easterly (representative of July condition) is more favorable for the amplification of Rossby-gravity (4, 0) mode with a theoretical period of ~2.5 days, while strong summer easterly (representative of January condition) favors the amplification of Rossby-gravity (3, 0) mode with a theoretical period of ~2.1 days. Rossby-gravity (4, 0) mode suffers a greater damping rate because of the lower group velocity, and therefore, it is more easily trapped in both latitude and altitude. These features of Rossby-gravity (4, 0) mode and the differences between (4, 0) mode and (3, 0) mode agree very well with our observational results of both W4 and W3.

4. Summary

[29] Coordinated global temperature and wind data sets during 2002–2012, observed by the SABER and TIDI instruments onboard the TIMED satellite, have been analyzed to study the seasonal and interannual variations of the W3 and W4. The two modes are evident in the wave number-frequency spectra of all the three components of temperature, meridional wind, and zonal wind. Our results show that the W3 maximizes primarily in January, while the W4 is amplified primarily in July and August. The W3 in temperature and zonal wind are mostly confined within the summer (southern) hemisphere, while the meridional wind component could extend from ~40°S in the summer hemisphere to ~30°N in the winter hemisphere. The W4 in both temperature and horizontal wind are confined within the summer (northern) hemisphere. The maximum amplitudes of W4 are ~8–9 K, ~40 m/s, and ~20 m/s in temperature, meridional wind, and zonal wind components, respectively, which are nearly half as large as those of W3, with ~15 K, ~65 m/s, and ~35 m/s.

[30] In January 2008 and 2009, the W3 was extremely weak with amplitudes of only ~5 K, 20 m/s, and less than 10 m/s, respectively, for temperature, meridional, and zonal components. Instead, abnormally strong W4 events are observed, which are as strong as the July/August W4 event. The weak summer easterly jets during these two years may not be strong enough to generate sufficient baroclinic/barotropic instability to amplify W3, but strong enough for the amplification of W4. This agrees well with the simulation result of Salby and Callaghan [2001] that relatively weak summer easterly is more favorable for the amplification of W4.

[31] It is also interesting to note the coincidence of the minima of W3 and the solar cycle minima around 2008 and 2009. We are not sure whether this is an evidence of a true geophysical connection between the amplitudes of W3 and the solar cycle or merely a coincidence. Extraordinarily enhanced W3 are observed during January 2006 in all the three components of temperature, meridional wind, and zonal wind, most likely related to the strong winter planetary wave activity occurring before a strong SSW event.

[32] Generally, the W4 mode starts to grow at late June and fades away before middle August with a lifetime of nearly 2 months, which is much longer than the life interval of 1 month for the W3. The W4 maximizes at ~70–75 km and 90 km in temperature and horizontal wind, lower than the peak heights of W3 at ~80–85 km and 92.5 km. The weaker magnitude values, lower peak heights, and longer life intervals of W4 suggest that W4 suffers a greater damping rate than W3. Our observations of W4 agree well with the features of Rossby-gravity (4, 0) mode, which is more easily trapped in both latitude and altitude because of the lower group velocity.

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