NOTES AND CORRESPONDENCE

Thermospheric Nocturnal Wind Climatology Observed by Fabry–Perot Interferometers over the Asia–Oceania Region

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Abstract

This study shows the horizontal structure of climatology of thermospheric nocturnal winds at a height of approximately 250 km in the Asia–Oceania region for the first time using observations made with Fabry–Perot interferometers (FPIs; optical wavelength of 630.0 nm). The FPIs used in this study were located at Shigaraki (Japan, 34.8°N, 136.1°E), Chiang Mai (Thailand, 18.8°N, 98.9°E), Kototabang (Indonesia, 0.2°S, 100.3°E), and Darwin (Australia, 12.4°S, 131.0°E). The observation data underwent quality control that involved consideration of cloud information, wind speed value, and standard deviation of results obtained from synchronous fringe images; approximately 30 % of the observation data from all the four stations were deemed suitable for use. The nocturnal diurnal changes at Shigaraki according to the local solar time were generally consistent with changes in China at similar latitudes, although the amplitudes were slightly different. All four stations showed the continuous flow pattern of the nocturnal diurnal wind in each season. The Chiang Mai and Darwin stations observed seasonal/diurnal changes similar to those observed by stations at similar latitudes on the North and the South American continents. Although there were fewer samples for Chiang Mai, Kototabang, and Darwin in the rainy season compared to that for Shigaraki, the seasonal climatology reported here can be used to provide a background long-term average status for describing anomalous events and extremes having different causes.

Keywords thermospheric wind; Fabry–Perot Interferometer; solar activity; ground-based observation

1. Introduction

Many factors affect the dynamics of the upper atmosphere, generally located 100–1000 km above

the sea level (a.s.l). The thermosphere–ionosphere coupled system is affected by solar extreme-ultraviolet heating, geomagnetic storms, and atmospheric oscillations propagating from the lower atmosphere. As illustrated in Supplement 1, the temperature profile in the thermosphere varies with the solar activity. Recently, the effect of solar activity on climate has become a growing concern, partly because of the issue of global warming (IPCC 2007;

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Russell et al. 2010). However, it is difficult to physically relate solar activity to the near-surface climate because there are many nonlinear processes involved (Gray et al. 2010). An understanding of the thermospheric physical processes is thus essential for explaining the possible links between solar activity and the lower and middle atmosphere.

The neutral wind is a primarily important parameter for understanding the dynamics and energy transport in the upper atmosphere. However, it is generally difficult to maintain a stable operation of the thermospheric wind (TW) measurement over a long period. The incoherent-scatter radar is able to estimate height-resolved meridional winds (Supplement 1; Wickwar et al. 1984; Witasse et al. 1998) but not zonal winds in the F region. Instruments onboard satellites and rockets are able to make instantaneous measurements. However, the Fabry–Perot interferometer (FPI), which can measure horizontal winds in the upper thermosphere (at approximately 250 km a.s.l.) based on Doppler shifts of the oxygen airglow emission at 630.0 nm (red line), is an instrument capable of long-term measurements.

The global TW climatology and its dependence on solar activities have been reported using United States-oriented FPI networks (Emmert et al. 2006a; Brum et al. 2012; Drob et al. 2015). Observations of TWs over the Asia–Oceania region have not yet been fully understood, except there have been recent studies in China (Yuan et al. 2013; Yu et al. 2014). The climatological TWs are dependent on the geographical latitude, season, and solar activity (e.g., Emmert et al. 2006a). In addition, the TWs may not be longitudinally symmetric; i.e., magnetic discrepancy between geomagnetic and geographic coordinates and nonmigrating tide can cause change in the TWs (Wu et al. 2014). One example of geographical-longitudinal dependency of TWs is the wave-4 structure observed in the equatorial ionosphere and thermosphere (e.g., Immel et al. 2006).

Our scientific goal is to depict the physics of the TW variation associated with solar activity. First, the background TW status of the Asia–Oceania sector must be presented. Here we report the processes of handling the FPI necessary for preparation prior to the climatological analysis. We also bridge the gap between the fields of meteorology and solar–terrestrial physics by showing observed thermospheric neutral wind with describing pre-process and handling processes.

2. Data availability and quality control (QC)

Observational data of the upper atmosphere are archived at individual research institutions (Yatagai et al. 2015). To make data more accessible, Emmert et al. (2006a) introduced the Coupling, Energetics and Dynamics of Atmospheric Regions (CEDAR) community archive and software with showing their FPI measurements and observed data. The Japanese solar–terrestrial physics community started the Inter-university Upper-atmosphere Global Observation NETwork (IUGONET) project in 2009 (Yatagai et al. 2014). The present study used FPI data obtained at Shigaraki, Chiang Mai, Kototabang, and Darwin, which are available from the IUGONET database (<http://www.iugonet.org>).

2.1 FPI observations

Table 1 summarizes coordinates of the stations used in this study. Details of the configurations of the instruments can be found at <http://stdb2.stelab.nagoya-u.ac.jp/omti/stations.html>. We used FPI data of red-line airglow (630.0 nm). The Doppler shift or the line-of-sight was calculated from individual fringes at four cardinal directions (i.e., north, south, east, and west in geographical coordinate) with elevation angle of 45°. The combination of data from the four cardinal directions provided two sets of horizontal components: eastward wind (V_e , meteorologically westerly, U) and northward wind (V_n , meteorologically southerly, V) (Shiokawa et al. 2003; Supplement 2).

2.2 QC and statistics of available data

FPIs cannot observe TWs if the sky is obscured by clouds. Even thin clouds may diverge Doppler shifts, which implies that the observed TW speed goes to 0 m s⁻¹ in the overcast. In addition, the intensity of airglow is important to obtain reliable observation. We therefore developed an objective QC scheme to eliminate abnormal values using information of the sky (cloud) status and the instability of estimated winds. We first describe the QC processes common to all four FPIs. The numbers of quality-controlled data are summarized in Table 1 and shown in Fig. 2.

a. Cloud status

We used hourly status reports of the sky condition, which are based on the manual verification of the cloud condition using an all-sky camera located at each station. These reports are available at <http://stdb2.stelab.nagoya-u.ac.jp/omti/obs1st.html>

Table 1. Location of the FPI instruments, observation periods, and sample numbers.

Station	Geo-graphic Latitude	Geo-graphic Longitude	Geo-magnetic Latitude	Geo-magnetic Longitude	Operational interval (period)	Local midnight in UTC	Number of samples (before QC; Ve, Vn)	Number of used samples (after QC; Ve, Vn)	Percentage of valid data (average of Ve and Vn)
Shigaraki (Japan)	34.8	136.1	25.4	205.1	2000–2014	About 15 UTC	63269, 63099	18794, 19171	30.1
Chiang Mai (Thailand)	18.8	98.9	8.9	171.5	2010–2014	About 17:20 UTC	14013, 14041	4616, 4991	34.2
Kototabang (Indonesia)	−0.2	100.3	−10.6	171.9	2010–2014	About 17:20 UTC	18780, 18780	4377, 4432	23.5
Darwin (Australia)	−12.4	131.0	−22.1	204.0	2011–2014	About 15:20 UTC	7958, 7996	2546, 2648	32.6

(or http://stdb2.stelab.nagoya-u.ac.jp/omti/obslst_cmu.html for Chiang Mai). The sky status is categorized as overcast or rainy (C), many clouds and a few stars (M), a few clouds and many stars (P), or clear sky with stars (S).

In our QC process, we decided to use data recorded under S and P status conditions. There were cases with no cloud information, which were not subjected to QC.

b. Abnormally large values

The physical behavior of TWs is not fully understood, and the plausible maximum wind speed is unknown. The present study regarded FPI-measured winds faster than the speed of sound in the thermosphere (750 m s^{-1} , see Supplement 1 of Balthazor and Bailey 2006) as abnormal values.

c. Threshold and status of each observation

1) Shigaraki

The FPI instrument installed at Shigaraki was described by Shiokawa et al. (2003). Horizontal winds are estimated from a pair of images, with each image having two fringes (cf. Supplement 2). The westerly component (Ve) is obtained from a pair of wind measurements (Doppler shift along the line-of-sight at opposite directions, cf. Supplement 2). A pair of wind estimates comprise two estimates made at an interval of 15 minutes, and for the estimation, we used both inner and outer fringes. Differences in Doppler shifts provide values of Ve_{inner} and Ve_{outer} . The values of Ve_{inner} and Ve_{outer} are basically estimated from the same pair of images and observed simultaneously. Thus, we take them as “one sample” after averaging Ve_{inner} and Ve_{outer} . The southerly (Vn) compo-

nents are observed and calculated similarly.

Figures 1a and 1b show the relationship between the photon count and the difference in velocity determined from the inner- and outer-fringe estimates. Weak airglow (i.e., a low count value of the charge-coupled device (CCD) camera) results in an unstable wind estimate and consequently large differences between the two fringe estimates ($|\Delta Vn|$); the same is true for $|\Delta Ve|$ as shown in Supplement 3.

It is revealed that the distribution after September 2010 (Fig. 1b) differed from that in earlier periods (Fig. 1a), which is attributed to the replacement of a cooling device of the CCD camera on August 30, 2010 (Personal communication with Shiokawa, June 2014, see acknowledgments). In both periods, $|\Delta Vn|$ takes a large value when the CCD count value is small, but notably small count values (< 100) were seldom observed after August 30, 2010. We therefore set different thresholds to discard weak airglow cases before/after the change in a cooling device. The threshold is a standard deviation (σ) of $|\Delta Vn|$ before (after) the change in cooling device of 53.5 (80.7). Therefore, we do not use the sample if the wind difference exceeds the threshold (σ of $|\Delta Vn| > 1$). The σ for $|\Delta Ve|$ are 55.8 (82.1).

2) Chiang Mai, Kototabang, and Darwin

The three FPIs adopted an optical design different from that of Shigaraki’s FPI (Shiokawa et al. 2012). They have 70 mm etalons, which are smaller than that of the FPI at Shigaraki, and they can estimate the wind component from 10 fringes in one image (cf. Supplement 2). One Vn sample is calculated from 10 north–south horizontal component pairs, and it is an average of up to 10 valid fringe values at one pair of

Fig. 1. a) Velocity difference (m s^{-1}) of the southerly component (V_n) plotted against the photon count value for Shigaraki before the exchange of the cooling device for the CCD camera (January 2000–August 2010). The vertical axis shows the absolute value of the V_n difference between the estimate from the inner fringes and that from the outer fringes ($|\Delta V_n| = |V_{n_inner} - V_{n_outer}|$). The dashed line and number indicate the threshold value. b) Same as a) but after the exchange of the cooling device (September 2010–December 2014). c) σ of one sample consisting of 10 pairs of westerly (V_e , m s^{-1}) estimates plotted against the photon count value for Chiang Mai. The dashed line and number indicate the threshold value. d) Same as c) but for southerly (V_n , m s^{-1}) estimates.

images. Instead of using the difference between two fringe estimates as for Shigaraki, the σ of more than five valid values is used to determine the threshold. Figure 1c shows the relationship between the σ within one V_e sample and mean count value within one V_e sample for Chiang Mai, while Fig. 1d shows the same for V_n . The same figures and statistics for Kototabang and Darwin are given in Supplement 4. The threshold ($\sigma > 1$) values of V_e (V_n) at Chiang Mai, Kototabang, and Darwin are 58.6 (44.5), 60.4 (60.6) and 34.1 (36.0), respectively. The figures clearly show that weak airglow causes a large σ within one sample.

2.3 Analysis method

The quality-controlled data are averaged for each sample time and for each component (V_e and V_n). All sample data are then placed into 15 min bins. We deal with the night-time diurnal variation of the TWs. Shigaraki has data for a period of 14 years and, for comparison with previous studies, we take the monthly averages of 15 min intervals. We apply a Gaussian-type filter to the climatological diurnal variation to obtain a smooth data. For the other stations, seasonal averages of 1 hour intervals (centered on the hour) are created. The seasons are boreal winter (November, December, January, and February), boreal summer (May, June, July, and August), and near equinox (March, April, September, and October).

2.4 Comparison of the overall average situation and quiet/disturbance cases

Most TW climatological studies eliminate cases of geomagnetic disturbance (i.e., $K_p > 3$ (Emmert et al. 2006a) or $Dst < -100$ (Yu et al. 2014)) to emphasize the response of the TW to geomagnetic disturbances (Emmert et al. 2006b). In contrast, our purpose is to obtain long-term mean values. Hence, for comparison with other climatologies of TWs, we present the difference between the overall average situation and quiet/disturbed cases at Shigaraki.

Figure 2a compares the overall average situation and cases of geomagnetic disturbance (storms) ($K_p \geq 3$). Although there are few storm cases (20.2 %), there is a large difference in the westerly component (V_e) from before to after midnight (15:00 UTC = 00:00 local time (LT)). The westward shift during the storm (i.e., geomagnetically disturbed) cases is often explained as a “disturbance dynamo” (Emmert et al. 2006b; Makela et al. 2014). In the quiet situation (Fig. 2b), V_e differs slightly ($\sim 20 \text{ m s}^{-1}$) from that of the overall average situation, whereas there is no significant difference in V_n . Therefore, only the differ-

ence in V_e (Fig. 2b) needs to be considered when we compare our results with the results of other studies.

3. Results and discussion

3.1 Climatology of Shigaraki

Pressure gradient is a major factor in controlling TWs at a height where the emission intensity of the red-line airglow has a peak. Therefore, TWs flow from the dayside to the nightside and from the summer hemisphere to the winter hemisphere. In the total averaged diurnal change (black bars in Fig. 2), the V_e component is a westerly before midnight, whereas it becomes an easterly from 02:00 LT. The delay in the change of direction at midnight is considered to be due to the time lag in the warming of the atmosphere and westerly flow. The V_n component is a southerly at dusk and dawn, but becomes a deep northerly at midnight. We then see the seasonal dependency of the timing, depth and magnitude of TWs.

Figure 3 shows the monthly mean values of V_e and V_n at Shigaraki. The general trend of V_e resembles that shown in Fig. 2. At equinox (March, September), the wind changes its direction at 2–3 hours after midnight (2–3 LT). This phase shift appears at a symmetrical point of the heated peak due to solar extreme-ultraviolet radiation, which peaks at approximately 15:00 LT (cf. Meriwether 2006). In winter, winds are mostly westerly, while in June and July, V_e turns easterly at midnight. V_n has a northerly component with a peak magnitude at midnight in all months, although the northerly amplitude is small from November to January. The northerly peak in some summer months (May, June, and August) appears 1–2 LT. Summertime’s strong northerly is probably due to the closer location to the dayside hemisphere (see Fig. 4).

Compared with observations made in China (Kelan, 37.7°N, 111.6°E), our V_e phases are similar to those of Yu et al. (2014; their Fig. 10) in all months, and this includes the westerly peak at approximately 12 UTC in November, December, and January. The phase and amplitude (0–80 m s^{-1}) is generally the same, although their winter amplitude is smaller than ours. Comparing our V_n with that of Yu et al. (2014; their Fig. 7), the northerly phase is similar in both, but the timing of the peak is sometimes different. Their amplitude in winter is larger than ours. These differences might be attributable to differences in the analysis year(s), because the observations in China were conducted for only 1 year.

We also compared our results with two FPI obser-

Diurnal Wind (630nm) Variation at Shigaraki

Fig. 2. Nighttime diurnal wind variation at Shigaraki. Black symbols show the overall average and number of samples (bar). Red symbols show cases of geomagnetic disturbance, defined as Kp index ≥ 3 . Blue symbols show geomagnetically quiet conditions, defined as Kp index < 3 . Bars indicate one σ away from either side of the mean values.

Ve & Vn Climatology at Shigaraki

Fig. 3. Nocturnal diurnal wind variation for each month at Shigaraki: (a) westerly (V_e) and (b) southerly (V_n). Thin lines are monthly averages and red lines are the Gaussian-filtered (temporally smoothed) changes in the relevant black lines. Local midnight and a wind speed of 0 m s^{-1} are marked with blue lines.

variations on the American continent (Emmert et al. 2006a) that latitudinally are closest to our sites: Millstone Hill (MH; 42.6°N, 71.5°W) and Arecibo (AC; 18.4°N, 66.8°W). In general, the overall features are similar, and V_e at MH (Emmert et al. 2006a; their Fig. 2) is closer than that at AC to that at Shigaraki. For V_n (our Fig. 3b; Emmert et al. 2006a; their Fig. 1), a peak northerly is observed before (after) midnight at AC (MH). Shigaraki has peaks at midnight through the year. Regarding the peak amplitude of V_n , MH has a stronger peak ($> 100 \text{ m s}^{-1}$) than AC and Shigaraki; however, the peak amplitude seems to be highly dependent on the solar activity or the data set analyzed.

TWs are affected by momentum transfer due to the particle-collisional process between the ionospheric plasmas and the thermospheric neutrals, and different behaviors can be seen in the plasma dynamics dependent on geomagnetic latitudes, as briefly mentioned in Section 1. Recently, Wu et al. (2014) compared simultaneous TW observations made in China and the United States (US) for two cases in October, and pointed out that V_n during low geomagnetic activities ($K_p \sim 2$) shows different characteristics before midnight (LT), although they show similar characteristics for V_e during low geomagnetic activities and for V_n and V_e during quiet geomagnetic condition ($K_p \sim 1$). The diurnal cycle of TWs over Asia–Oceania may differ from that in the Western Hemisphere, and the relationship should depend on the geomagnetic conditions. It is expected that long-term observations will be conducted to clarify the physical mechanism. In any case, as the geomagnetic latitude affects the TW climatology, our Asia–Oceania TW climatology described here is needed to highlight various “cases” or “extremes” that occur in the upper atmosphere over Asia–Oceania.

3.2 Climatology of other stations

Here we present hourly averaged results for the three stations.

a. Chiang Mai (Thailand)

Figures 4a–c show the seasonal (four-month) averaged TWs at Chiang Mai. At this latitude, the airglow intensity is stronger than that at mid-latitudes, and more valid data were obtained than at Shigaraki, especially, in wintertime and before midnight. V_e is always westerly and V_n has a midnight peak of southward wind. During equinox and boreal summer, the peak appears earlier by 1–2 hours.

This pattern is similar to that of MH in winter for

a high solar activity case in terms of peak amplitudes and timing. The negative V_n before midnight is a characteristic similar to the described for the equinox and summer cases at AC (Emmert et al. 2006a). In summer, Chiang Mai experiences a wet monsoon, which results in a small sample number and large σ .

b. Kototabang (Indonesia)

Figures 4d–f show the results for Kototabang. This site is located near the equator (0.2°S). The σ of V_n are very small. V_e is generally a weak westerly throughout the year, while V_n is a weak southerly (northerly) in boreal winter (summer) before midnight. Even the change in V_e in each season is small, and the pattern is very similar to that for Brazil (Meriwether et al. 2011, their Fig. 4). Care should be taken for values with large σ and a small sample number (our Fig. 4e, 12:00–16:00 UT), but the general trend is considered to be true. A longer observational archive is needed to address further characteristics.

c. Darwin (Australia)

Figures 4g–i present the results for Darwin, Australia. We compare the tendencies of TWs with those of Brazil (latitude 6.89°S), where 1-year averaged TWs are displayed (Meriwether et al. 2011; their Figs. 3, 4). The diurnal variations of TWs in Australia and Brazil resemble each other for all three seasons. In winter, for example, there is a peak around 12:00 UTC (21:00 LT) in both V_e and V_n , and the subsequent tendency of the values is to become small. Near equinox and during austral winter, V_e appears similar to that in Brazil, but the amplitude is smaller than that in Brazil. The amplitude of V_n in austral winter is low, but the phase is similar to those in May, July, and August of 2010 for Brazil (Meriwether et al. 2011; their Fig. 3). Since we also used data obtained in 2010, the similarity may be attributed to having used data partly from the same period, and smaller amplitude may be because we averaged more data.

3.3 General view

Figure 5 shows summary plots for the four stations in each season. For Shigaraki, we took the same 4-month averages as for the other three stations. A clear seasonal difference is observed, and the wind patterns observed by the four stations show a consistent structure. The general pattern near equinox is consistent with the tendency of thermospheric general circulation under the solar-minimum equinox condition (Dickinson et al. 1984), which has a center of

Fig. 4. (a)–(c) Nocturnal diurnal wind variation at Chiang Mai, Thailand: (a) boreal winter, (b) near equinox, and (c) boreal summer. The sample number is based on V_e . Bars indicate one σ away from either side of the mean values. Blue lines indicate midnight estimated from the geographic longitude (see Table 1). (d)–(f) Same as (a)–(c) but for Kototabang, Indonesia. (g)–(i) Same as (a)–(c) but for Darwin, Australia.

Thermospheric Wind Pattern

Fig. 5. Summary of wind images for each season. Each small globe shows the night sphere (solstice or equinox) and location of the four stations. The base of each arrow on the large globes is the geographic latitude of one of the four stations, and longitude does not correspond to the map but gives the local time (LT). The direction of each arrow corresponds to the map.

convergence (minimum temperature) at 04:00 LT for solar heating only (their Fig. 8a) and solar heating and geomagnetic convection (their Fig. 8b).

In boreal summer, a deep northerly component through the night and an easterly component after midnight are dominant. There are asymmetric features in the comparison of winter and summer, as we do not have large numbers of data for the Southern Hemisphere. Therefore, additional observation data are required, especially during years of high solar activity.

4. Summary

According to four FPI observational data sets obtained in the Asia–Oceania region, we presented the climatology of the nocturnal neutral wind of the thermosphere (TW). After QC, about 30.1 % of observation samples were judged as being valid data (Table 1). It is sufficient to discuss the monthly state of TWs for the Shigaraki station and the seasonal state of TWs for the three other stations. We found that the diurnal changes at Shigaraki were generally consistent with those observed in China at similar latitudes, although the amplitudes were a little different. The other three stations showed seasonal differences consistent with those observed at Shigaraki, and the Chiang Mai and Darwin stations observed seasonal/diurnal variations similar to those observed by stations at similar latitudes on the American continent. These monthly/seasonal climatological diurnal TW data should be used to represent the background structure when investigating specific events in the thermosphere.

Supplement

Supplement 1 shows general features and diagnostic instruments of the upper atmosphere. Supplement 2 shows schematic diagrams of FPI observations and data processing processes.

Supplement 3 shows the relationship between the photon count and the difference in velocity at Shigaraki.

Supplement 4 shows the relationship between the standard deviation within one sample and the mean count value within one sample for Chiang Mai, Kototabang, and Darwin.

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