Resolving Vertical Variations of Horizontal Neutral Winds in Earth's High Latitude Space-Atmosphere Interaction Region (SAIR)

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Abstract  Few remote sensing or in-situ techniques can measure winds in Earth's thermosphere between altitudes of 120 and 200 km. One possible approach within this region uses Doppler spectroscopy of the optical emission from atomic oxygen at 558 nm, although historical approaches have been hindered in the auroral zone because the emission altitude varies dramatically, both across the sky and over time, as a result of changing characteristic energy of auroral precipitation. Thus, a new approach is presented that instead uses this variation as an advantage, to resolve height profiles of the horizontal wind. Emission heights are estimated using the Doppler temperature derived from the 558 nm emission. During periods when the resulting estimates span a wide enough height interval, it is possible to use low order polynomial functions of altitude to model the Doppler shifts observed across the sky and over time, and thus reconstruct height profiles of the horizontal wind components. The technique introduced here is shown to work well provided there are no strong horizontal gradients in the wind field. Conditions satisfying these caveats do occur frequently and the resulting wind profiles validate well when compared to absolute in-situ wind measurements from a rocket-borne chemical release. While both the optical and chemical tracer techniques agreed with each other, they did not agree with the HWM-14 horizontal wind model. Applying this technique to wind measurements near the geomagnetic cusp footprint indicated that cusp-region forcing did not penetrate to atmospheric heights of 240 km or lower.

Plain Language Summary  We present a fundamentally new capability for measuring height variations in horizontal winds for altitudes between 100 and 150 km. It is difficult to measure these winds because space-based platforms are unable to maintain such low orbits, whereas ground-based remote sensing depends on an optical emission that behaves in ways that are difficult to account for. In particular, there are times when the emission height varies dramatically which, historically, made derivation of horizontal winds intractable. Here, we instead use this variation as a means to resolve how the wind changes with height. Not only does this allow height profiling, but we can now also make use of previously rejected data periods. Our results were validated by comparing to absolute wind measurements from a chemical tracer released by a sounding rocket. Comparing these profiles to the most-commonly used empirical wind model (HWM14) showed that the model does not agree with the measured wind behavior. Finally, we have applied this technique to one of Earth's least understood atmospheric regions: the terrestrial footprints of the polar cusps. Results of this study placed a lower bound on the altitudes influenced by the cusps; its effects were not observed below 240 km.

1. Introduction

Optical remote sensing studies of thermospheric winds during auroral activity have focused mostly on Doppler spectroscopy of the 630 nm atomic oxygen airglow/auroral emission (e.g., Billett et al., 2020; Conde et al., 2018). The 630 nm red line emission's height distribution and spectral characteristics make it a simple and reliable indicator of middle thermospheric (F-region) conditions (Smith et al., 1988). The lower thermosphere, especially from about 120 to 200 km (E-region), is more difficult to study because there are few, if any, simple diagnostic methods that can produce accurate long-duration observations of winds in this height range.

Space-based in-situ measurements such as the Wind and Temperature Spectrometer (WATS; Spencer et al., 1981) instrument aboard the Dynamics Explorer 2 (DE-2) spacecraft (Killeen & Roble, 1988) measured the angle of arrival of the apparent wind to determine cross-track wind components between approximately 300 and 650 km altitude (e.g., Innis & Conde, 2002). DE-2 also used a Fabry-Perot Interferometer (FPI) to measure...
the along-track component of the wind (Hays et al., 1981). At the lower altitudes of interest here, orbital lifetimes are very short. Aerodynamic drag increases rapidly below 300 km altitude, making long-duration in-situ measurements of Earth's E-region impossible with current platforms (e.g., Bock et al., 2011). Remote sensing from space-based instruments can resolve height variations of thermospheric winds through limb viewing. For example, the wind imaging interferometer (WINDII) aboard the Upper Atmosphere Research Satellite (URS) used a Michelson interferometer to make measurements of visible airglow emissions in the limb from 80 to 300 km altitude. Horizontal winds were derived from Doppler shift measurements (Shepherd et al., 1993). Such techniques can provide global-scale spatial measurements. However, they fail to provide temporal resolution at an individual location due to the spacecraft's constantly moving observing area. Without sufficient temporal and spatial coverage, studies such as the response of the thermosphere to changes in auroral forcing are difficult at best. Sounding rockets performing chemical releases can provide infrequent, short periods of wind measurements in the region between 80 and 300 km (e.g., Larsen, 2013; Zhan, 2007). Chemicals such as trimethyl aluminum (TMA), lithium, sodium, barium, or strontium are released from sounding rockets to form a trail or discrete clouds along the rocket's trajectory. Height profiles of thermospheric winds can be derived using triangulation techniques. The motion of the tracers in three-dimensional space is determined by photographing the tracers over time from several separate locations and using geometric triangulation to determine their location in three-dimensional space. This technique is expensive, chemicals diffuse away quickly (limiting temporal coverage), and profiles are spatially limited to local measurements near the rocket trajectory. Therefore, chemical releases are also not a viable large-scale or long-term measurement technique. However, the technique does provide absolute measurement of winds in the region, due to the small sampling volume and lack of assumptions or separate calibrations needed for accurate triangulation. Camera look angles are absolutely calibrated by using stars appearing in the image background as reference points. Indeed, chemical releases are the only technique currently available that can provide absolute measurements of thermospheric winds.

Ground-based active techniques to measure thermospheric winds typically only provide reliable measurements up to ~120 km altitude. Light Detection and Ranging (LIDAR) systems use Rayleigh scattering to measure Doppler shifts (e.g., Alpers et al., 2004) or use resonance fluorescence techniques that depend on the presence of metals in the atmosphere. Fluorescence LIDAR system returns have been obtained under favorable conditions from as high as 170 km (e.g., Gao et al., 2015), although reliable routine wind measurements have not yet been demonstrated from this altitude. However, LIDAR systems have for some time been able to resolve winds up to 110 km altitude (e.g., Dao et al., 1995). Recently this technique has been used by the Andes Lidar Observatory to measure winds at 140 km, although literature on the reliability, resolution, and results of these measurements is still very limited (A. Z. Liu et al., 2016). Various radar techniques exist to measure winds over a range of altitudes. These include meteor radars (e.g., Wilhelm et al., 2017), medium-frequency (MF) radars (e.g., Phillips et al., 1994), and incoherent scatter radars (Nicolls et al., 2007). Meteor radars can be reliable indicators of the winds up to about 100 km, whereas incoherent scatter radars can resolve winds up to 120 km (e.g., Heinselman & Nicolls, 2008; Oyama et al., 2009). MF radars can be used to measure winds to perhaps as high as 110 km. However, above 90 km altitude, the reliability of measurements decreases drastically (e.g., Vincent et al., 1998). High altitude balloons for measuring winds are unable to reach the altitudes of interest as they only fly below 50 km. One of the most common ground-based techniques is to use the high spectral resolution afforded by FPIs to measure Doppler shifts from airglow emissions. From Doppler spectral widths and Doppler shift measurements, temperature and wind estimates can be derived (e.g., Anderson, 2011).

Despite being difficult to measure, the region between 100 and 200 km altitude is important to understand for several reasons. It is the transition between atmosphere-like and space-like environments. Many fundamental changes in the atmosphere happen in this region: for example, absorption of solar ultraviolet radiation, dissipation of semi-diurnal tides that propagate up from below, and the transition from ion flows being driven by the background neutral wind at lower altitudes to the E x B drift at higher altitudes (e.g., Sangalli et al., 2009). These can have significant effects on technologies such as communication, navigation, radar tracking, and spacecraft orbiting just above this region. Many observations from sounding rocket experiments over the years have shown that there exist strong vertical gradients in the horizontal wind, specifically, between about 100 and 150 km altitude (e.g., Larsen, 2002; Larsen & Fesen, 2009). Thus, there is considerable operational interest in the vertical gradients that occur in this region. Above 150 km, the vertical gradients in the winds are commonly assumed to be small enough that measurements from 240 km and above can be used to infer winds down to the base of...
The aurorally excited 558 nm atomic oxygen emission occurs within the same altitude region as the strong vertical gradients in the winds discussed above. Further, the altitude of the aurorally excited emission varies as a function of space and time. The height profile of the aurorally excited 558 nm emission is determined by the depth to which precipitating auroral electrons can penetrate into the atmosphere (Lummerzheim & Lilensten, 1994). This depth depends, in turn, on the characteristic energy of electron precipitation. It is possible to study the variation of this characteristic energy and resulting height variations in the 558 nm emission layer using Scanning Doppler Imager (SDI) instruments (e.g., Holmes et al., 2005; Kaeppler et al., 2015). The SDIs are ground-based all-sky imaging Fabry-Perot spectrometers that were first introduced by Conde and Smith (1997). SDI observations from Alaska of the 558 nm Doppler temperature indicate that the characteristic energy can often be relatively constant with respect to both time and space. However, there are many scenarios, such as the substorm growth phase, when the auroral characteristic energy varies considerably between different look directions in the sky and over time (Holmes et al., 2009). Emissions observed during these periods, therefore, can originate from anywhere within a wide range of altitudes, that is, roughly 100–150 km.

Fabry-Perot spectrometers have been used for decades to estimate winds based on Doppler shift measurements such as those obtained from the 630 nm emission (e.g., Burnside et al., 1981; Yagi & Dyson, 1985). However, these methods to construct wind fields from Doppler shift measurements in multiple look directions do not typically account for situations in which different altitudes are sampled. Such altitude variations would be inconsequential if there was little variation in the wind field with altitude. However, TMA chemical release experiments show that vertical shear in the wind field is common in this region (e.g., Larsen, 2002), and therefore observing and analysis techniques that do not specifically account for those shears are likely to produce erroneous wind fields. The purpose of the present work is to instead use this as an advantage, by exploiting the altitude variations of the emission over time and space to infer a height profile for the winds. Then, this method can be used to measure winds in the 100–150 km region, at least during periods when spatial and temporal variation of the emission altitudes spans a sufficiently extended range of heights. When emission heights are fairly uniform, as indicated by a uniform temperature across the sky, useful wind reconstructions are possible using the same monostatic wind analysis that is routinely applied to the 630 nm emission data to construct a wind field (Conde & Smith, 1998).

### 2. Observational Techniques

The present work is based on data from SDI instruments. The specific instruments used in this study and their geographic locations can be seen in Table 1. The SDI field of view is upward looking, zenith centered, and typically spans a half angle of 70°–75°. A fish eye lens and additional coupling optics map this very wide object-space field of view onto the much less divergent beam that must pass through the etalon. The field of view is mapped such that the fringes from the Fabry-Perot etalon are conjugate with the sky. The etalon gap is scanned periodically through a distance corresponding to one order of interference at the observing wavelength, which means each pixel records a complete spectrum of the emission line spanning one free spectral range of the etalon in the wavelength dimension. However, the nature of the Fabry-Perot ring pattern is such that the spectral response of pixels mapping to different radii from the center of the ring pattern is not in phase with respect to the variation of etalon gap. The image processing software must, therefore, apply a real-time phase shift as the etalon scans to bring the responses of all pixels into a common spectral phase. This allows the image processor to co-add signals from groups of pixels in any arbitrary configuration to increase the signal-to-noise ratio of the accumulated spectra at the expense of reduced angular resolution in the sky. Within the field of view, the sky is divided into sections called zones, where each zone of the sky will have an independently measured and recorded Doppler spectrum of the emission line. Figure 1a shows SDI data from the 558 nm spectrum observed from Toolik Lake, Alaska (Table 1). The emission line observed can be changed by simply changing to a different filter in the system and adjusting the scan range for the etalon gap appropriately. By numerically fitting modeled

| Instrument                  | Longitude | Latitude |
|-----------------------------|-----------|----------|
| Poker Flat, Alaska          | −147.430  | 65.1192  |
| Toolik Lake, Alaska         | −149.60   | 68.633   |
| McMurdo Station, Antarctica | 166.666   | −77.850  |
| South Pole Station, Antarctica | —        | −89.999  |
spectra to the sky spectra, a Doppler shift and Doppler width for each zone can be derived. This technique has been previously used and is well described by Conde et al. (2018), Conde and Smith (1997), Anderson (2011), Anderson et al. (2009), Dhadly and Conde (2017), and Dhadly et al. (2015). The Doppler shift and Doppler width derived from the SDI spectra measurements are then used as input data to the algorithm described here, which creates the height profile.

When the characteristic energy of the auroral electron precipitation increases, it allows the particles to penetrate deeper into the thermosphere. There they excite optical emissions from a cooler population of radiating atoms and hence exhibit a lower Doppler temperature. This is well demonstrated in Figure 1b. In the figure, it is apparent that the colder region is approximately co-located with the area of bright aurora. The colder 558 nm Doppler temperatures are assumed in this work to originate from lower altitudes. As discussed previously, this implies that the characteristic energy of the precipitating auroral electrons producing the bright arc was higher than the precipitation further equatorward. In order to exploit this effect, the MSIS atmosphere model (Figure 2) with the appropriate F10.7 and Ap values for the night is used to map the measured Doppler temperature to the corresponding centroid emission height. That is, it is possible to look up an emission altitude by “inverting” the model results such that height is treated as a function of temperature. From this, it is straightforward to use numerical interpolation to predict the emission height that corresponds to each temperature derived from the observed 558 nm Doppler spectra. Assumptions and possible problems with this method and using MSIS will be addressed in Section 5, titled “Caveats.”

Height profiles of horizontal wind vectors are generated using the line-of-sight wind estimates derived from SDI spectral data. The time periods chosen for analysis consist of up to several tens of “exposures,” each of which includes line-of-sight winds from 115 look directions. These measurements are reorganized into a one-dimensional vector of $N$ scalar values. For each line-of-sight measurement there is an associated zenith angle, azimuth angle, and altitude. First, a mathematical model must be formulated to describe the wind as a function of height. The wind is separated into two components:
zonal and meridional. Both of these components are modeled as varying with height but constant in the model with regard to horizontal displacements. The model, \( \mathbf{M} \), fits a separate seventh order polynomial to represent each of the zonal and meridional components of the wind such that

\[
\mathbf{M} = U(z)\hat{x} + V(z)\hat{y}
\]

where \( U(z)\hat{x} \) represents the zonal component and \( V(z)\hat{y} \) represents the meridional component. The polynomials are represented as

\[
U_i = u_0 + u_1(z_i - z_0) + u_2(z_i - z_0)^2 + \ldots \\
V_i = v_0 + v_1(z_i - z_0) + v_2(z_i - z_0)^2 + \ldots
\]

and the index \( i \) refers to the \( i \)th line-of-sight wind measurement, which will be denoted as \( L_i \). The variable \( z_i \) is the altitude inferred for the \( i \)th wind measurement based on the Doppler temperature, and \( z_0 \) is a convenient reference height used to center the polynomial expansion. In this case, \( z_0 \) was chosen to be 120 km. We model the \( i \)th line-of-sight wind measurement as

\[
L_i = U(z_i) \sin \theta_i \sin \phi_i + V(z_i) \sin \theta_i \cos \phi_i.
\]

where \( \phi_i \) is the azimuth angle measured eastward from magnetic north for observation \( i \) and \( \theta_i \) is the zenith angle for observation \( i \).

Substituting Equations 2 and 3 into Equation 4 gives

\[
L_i = \left( u_0 + u_1(z_i - z_0) + u_2(z_i - z_0)^2 + \ldots \right) \sin \theta_i \sin \phi_i \\
+ \left( v_0 + v_1(z_i - z_0) + v_2(z_i - z_0)^2 + \ldots \right) \sin \theta_i \cos \phi_i
\]

or

\[
L_i = u_0\sin \theta_i \sin \phi_i + u_1(z_i - z_0) \sin \theta_i \sin \phi_i + u_2(z_i - z_0)^2 \sin \theta_i \sin \phi_i + \ldots \\
+ v_0\sin \theta_i \cos \phi_i + v_1(z_i - z_0) \sin \theta_i \cos \phi_i + v_2(z_i - z_0)^2 \sin \theta_i \cos \phi_i + \ldots
\]

simplifying to

\[
L_i = \sum_{j=0}^{14} u_j \sin \theta_i \sin \phi_i(z_i - z_0)^j + \sum_{j=0}^{14} v_j \sin \theta_i \cos \phi_i(z_i - z_0)^j.
\]

This can be considered as a sum of 14 basis functions, each of which can be regarded as a function of a single independent variable, \( i \), multiplied by its corresponding \( u_j \) or \( v_j \) coefficient. These 14 coefficients must be determined such that the \( \chi^2 \) sum of deviations is minimized with the observations \((O_i)\) where

\[
\chi^2 = \sum_i (L_i - O_i)^2.
\]

Writing the complete set of coefficients as

\[
\{ b \} = \{ u \} \cup \{ v \}
\]

we then require that

\[
\frac{\partial \chi^2}{\partial b_j} = 0.
\]

Since the coefficients, \( b_j \), appear linearly as seen in Equation 7, the derivative gives a system of linear equations. To solve, we use the singular value decomposition (SVD) technique implemented by IDL. This process is discussed more in depth in Golub and Reinsch (1971).
Figure 3. (a) Routinely generated survey plot summarizing the neutral wind components inferred from data recorded by the Scanning Doppler Imager at Poker Flat, Alaska on the night of February 2, 2003, using a conventional technique that takes no account of height variations. At every time step, there are estimates of both the zonal and meridional winds for all zones. The solid curves on the top two panels indicate the median winds calculated across all zones at each time step. Error bars denote the two sigma standard deviation across all zones at each observation time. The bottom panel shows the line-of-sight wind component derived from the zenith zone (green). The pink line indicates the vorticity calculated from the horizontal wind components across all zones at each time. Derived winds between 6 and 10 UT exhibit the noisy values that often indicate favorable candidate periods for height profiling. (b) Routinely generated survey plot of the median Doppler temperatures computed across all zones for each time, as inferred from the SDI data at Poker Flat, Alaska on the night of February 2, 2003. Red contours show the MSIS altitude. Note that the observed temperatures correspond to varying MSIS altitudes over time.
The observations depend on the known variables of time, zenith angle, and azimuth angle. Once the polynomial coefficients are solved, the only variable left in the model is altitude. The SVD algorithm only allows for one input variable. To numerically fit the data as described above, observations are stored in a 1D array where the array index, \( i \), points to an SDI observation in a specific zone with unique time, zenith angle, and azimuth angle values. This allows for all dependencies to be written in terms of the single variable \( i \) which can then be passed to the fitting routine.

3. Data

Obviously, it is only possible to derive height profiles from time periods during which the 558 nm emission altitude was varying, since the fitting procedure needs height variation to constrain the best values for the modeled coefficients. Candidate periods are identified by scanning through automatically generated survey plots. Two indicators that occur in the survey plots are periods when the vector wind components fitted to the 558 nm Doppler shift data look noisy, or when the inferred 558 nm Doppler temperature varies. In Figure 3a, the noisy wind field seen between 6 and 10 UT likely indicates routine wind fits were degraded by rapidly changing emission altitudes. In order to create a height profile for a period of time, the emission height must be varying both in time and across the sky, with the former variation being required to ensure a range of heights is observed at each azimuth and the latter variation being required to allow the routine to distinguish between the meridional and zonal components.

Figure 3b shows the variation of temperature over time for this night. Here, the temperature is derived by taking the median of the Doppler temperatures over all look directions for each exposure with the result plotted as a function of the central time of the exposure. This plot is automatically generated as a quick look indicator for each night. More temperature variations than those seen in this figure are observed due to the median sampling concealing detailed variations. Since the temperature of the emissions provides an effective proxy for the emission altitude, this figure effectively indicates height variations averaged over the sky. In this plot, there are large and sudden changes in temperature. A swift increase in temperature occurred at around 6 UT. The temperature
was variable but declining until around 10 UT which corresponds well with the noisy period in the zonal and meridional components of the wind from Figure 3a. Red contours in Figure 3b show the MSIS model height profile for the night. This further shows that increases in temperature correspond to increases in altitude.

Now the focus is restricted to the time period from 08:00:28 to 09:00:49 UT when Doppler temperature was varying. Figure 4 shows the Doppler temperature plotted for each SDI exposure over the course of the night. From this plot, it is apparent that the temperatures at the beginning and end of the night were fairly consistent across the field of view and over time. Therefore, during those time periods, wind reconstruction techniques that assume a single emission altitude when combining data from different locations in the sky could operate reliably.

Figure 5a shows Doppler temperatures during exposures from the restricted time, and Figure 5b shows the inferred altitude of emissions over the same time period. As discussed before, the altitude plot was generated using the MSIS model (depicted in Figure 2) to look up the altitude for each zone given the measured Doppler temperatures. Comparison of Figures 5a and 5b shows the relation between temperature and altitude of emissions.

Finally, the height profiling technique is applied to this time period to resolve the wind profile with altitude. The code generates target maps (Figure 6a) as well as the height profile (Figure 6b). The target map in Figure 6a shows how all measurements from the selected time period map in reference to the SDI (center of target) as if viewed from above the observations. Each observation is plotted as a colored mark on the target map. The color of the

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**Figure 5.** (a) Every Scanning Doppler Imager exposure during the period of 08:00:28 to 09:00:49 UT. The Doppler temperature variation is shown in blue to red hues as indicated in the legend. SDI exposure times typically were 1 min 30 s but can vary by up to 35 s. Central times of each exposure can be approximated by adding the appropriate multiple of 1 min 30 s to the start time of 08:00:28 UT. Although this is only an approximation, departures from the actual observing times have little geophysical significance because the time scale of changes is much longer than the observed cadence depicted here. (b) SDI exposures during the period of interest showing the calculated MSIS altitude of emission in blue to red hues.
mark is determined by the fitted altitude as indicated by the color bar. The location of each mark is determined by calculating the appropriate radial distance using the azimuth, elevation, and height for the zone it was measured in. Emissions from higher altitudes (red) will always map to a larger radial distances from the SDI within the same zones due to the cone-shaped field of view. Ideally, target maps would include observations from all altitudes in all azimuths.

Height profiles of the zonal and meridional wind components, as depicted in Figure 6b, are the main product of this process. On the left side of the figure, there is a histogram showing the number of spectra observed within a set of altitude bins. In the center of the plot are the meridional and zonal wind fits. This also demonstrates how substantially the wind velocity can vary with altitude, which further emphasizes the need for this technique. Zonal wind speeds near 110 km altitude were approximately −75 m/s and increased to −190 m/s just 30 km higher.

4. Results

To validate the height profiling method, comparisons are made with winds inferred from the drifts of chemical tracer clouds released from a sounding rocket. Although sounding rocket flights provide infrequent measurements, each of which span short time intervals, the data collected are essentially an absolute measurement. This is because the tracer must move with the wind due to the relatively high collision frequency between the tracer molecules and the ambient atmosphere. (Based on MSIS atmospheric profiles, this frequency varies very roughly from ~100 Hz at 120 km altitude down to ~10 Hz at 150 km). Further, at high altitudes, the high molecular weight tracers are known to sink through the background atmosphere. However, even at the highest altitudes of interest here, the sedimentation speed is slow compared to both the horizontal and vertical wind speeds (Rieger, 1974). The geometry of the triangulation is absolutely referenced with respect to the distant stars in the sky images and requires no further a priori assumptions. Furthermore, of the tracers commonly used, trimethyl aluminum (TMA) is appropriate over the specific height range of interest (100–150 km altitude) because it is easily visible for at least several minutes, and can thus be tracked for long enough to characterize the wind. Poker Flat Research Range hosts an SDI, and simultaneously it is one of the locations from which rocket-based chemical release measurements of thermospheric winds have occurred.

Figure 6. (a) Target map generated by the height profiling technique showing the altitude coverage locations, as described in the text. The observing site is located at the center of the plot. Blue to red hues show the altitude inferred for each spectrum recorded. (b) The 1D height profiles for zonal (positive for eastward and negative for westward) and meridional (positive for northward and negative for southward) winds associated with the target plot shown in panel (a) The histogram at the left shows the number of data points at the emission altitude.
During many of the TMA rocket launches from Poker Flat, the Alaskan SDI network was only recording 630 nm spectra. However, for the March 27, 2003 “JOULE” rocket mission, the instrument recorded 558 nm auroral emission data allowing for comparison of the height profiling technique to the winds derived from triangulation of the TMA released from the JOULE chemical release rocket.

Figures 7a and 7b show the temperature and related heights derived from the SDI spectra over the JOULE launch period, respectively. Notice that around the time of the downward leg of the rocket trajectory (12:13 UT), the 558 nm emission height variation was minimal compared to the previous example of this method (Figure 5). Figure 8a, the generated target map, further shows that there were few instances of 558 nm emissions from higher altitudes. However, of the few measurements from higher altitudes, they are spread over the field of view rather

Figure 7. (a) Time evolving sky maps of Doppler temperature (blue to red hues) derived from the Poker Flat Scanning Doppler Imager on the night of the JOULE sounding rocket launch on March 27, 2003. The first exposure was taken at 11:45:10 UT and the last exposure at 13:01:22 UT with an average exposure time of 1 min 30 s and the largest deviation being 22 s. Similarly to Figure 3, central exposure times can be approximated by adding the appropriate multiple of the average exposure time to the start time. (b) Maps of calculated MSIS height (blue to red hues) over the same time period. Note that similarities in Doppler temperature and height distributions occur because temperatures from (a) are used to assign altitudes in (b).
than clustered together. Figure 8b shows the derived height profiles. The histogram in Figure 8b also reflects the relative lack in higher altitude measurements. The majority of spectra originated from altitudes between 115 and 130 km. The number of observations above 130 km was relatively low when compared with lower altitudes. However, the histogram scale on the top of Figure 8b shows there were still tens of observations per bin in those higher altitudes. This is a more typical example of how the spread of data tends to appear. The overall hundreds of data points at higher altitudes still provided enough coverage for fitting two seventh order polynomials.

Upon examination of these profiles, they agree well with the data recorded by the JOULE mission as shown in Figure 9. Figure 9a shows a close-up of the region of data covered by the height profiling technique and Figure 9b shows the full profile from the rocket TMA trail. As expected, the 1D height profiling technique was not able to resolve the small-scale structures in altitude that are present in the rocket profile. However, regardless of the possible lack of detailed height variation, the agreement between the SDI data and the absolute in-situ wind measurements is encouraging.

This method was able to retrieve a reasonable height profile for the night of the JOULE launch. It can also be shown that the wind fitting is stable on this night. By experimenting with including and excluding subsets (in time) of the data during this period, it was found that the fitted result is largely insensitive to exact choice of exposures used. An example of this process is shown in Figure 10, where results derived during the restricted time period still correspond closely to actual data from the launch.

5. Caveats

The method described has been applied to many days of data. Resulting height profiles were stable with respect to the choice of spectra used to calculate the wind profile and generally consistent with expectations based on horizontal wind fields derived from the traditional 2D fitting technique, as well as the range of altitude variations seen in rocket data. However, in some instances, the derived wind profiles were clearly incorrect.

Because the 1D height profiling technique only fits altitude variations, it does not have a mechanism to account for horizontal gradients in the wind field. Figure 11a shows a horizontally uniform wind field that resulted from traditional analysis of 558 nm spectra. Figure 11b shows an example of a horizontally nonuniform wind field generated by the traditional methods. During both periods, the emission altitude varied vertically and horizon-

Figure 8. (a) Generated target map from the JOULE launch period. (b) Height profiles of the JOULE launch period. Altitude coverage appears sparse above 130 km in (a) and (b) but, as noted in the text, the actual number of observations is sufficient.
tally as indicated by the Doppler temperatures. The 1D height profiling technique uses the azimuthal variation of the line-of-sight wind component to distinguish between meridional and zonal winds. It is formulated under the assumption that every azimuth angle is sampling the same horizontally uniform wind such that the line-of-sight component would exhibit a single cycle of a harmonic oscillation over the full 360° azimuth. As shown in Figure 11c, a good match is usually obtained when fitting a single-cycle harmonic oscillation to the line-of-sight component of a horizontally uniform wind field (Figure 11a). During time periods when the wind field is horizontally sheared, as in Figure 11b, the line-of-sight component of the observed winds typically exhibits two cycles of oscillations over 360° of azimuth. The fitting of a single-cycle harmonic oscillation to the observed double-cycle harmonic oscillation is shown in Figure 11d. This results in a very small fitted amplitude for the harmonic oscillation since it is fitting a basis function with a primary form that is orthogonal to that of the data. Therefore, before using a period of time that is potentially acceptable based on altitude variations, the routinely generated 2D horizontal wind field must be checked for horizontal uniformity. Using our standard Fourier fitting technique (Conde & Smith, 1998) to generate a 2D wind field for a candidate time period that is viable for the 1D height profiling technique will not provide a truly accurate representation of the wind field because, by definition, it will involve merging observations originating from different altitudes. However, it is adequate for use simply as a check for horizontal uniformity, because it is very unlikely that artifacts due to height variations would align exactly enough to make a nonuniform wind appear uniform.

This study uses the MSIS model to assign altitudes to the line-of-sight wind measurements. This model was first introduced in 1977 (Hedin et al., 1977) and has been revised a number of times since. The version used here is MSIS-90. While this is not the latest version, the temperature behavior over the height range of interest here has not changed substantially enough to impact our analysis in the more recent instances, and we have continued using the earlier version for historical reasons. Further, we have compared 630 nm SDI Doppler temperature data with

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**Figure 9.** Comparison of zonal and meridional wind components derived from the JOULE rocket TMA release starting at 12:13 UT (red) with those obtained from the height profiling technique (black). Panel (a) emphasizes the height range spanned by the optical technique and shows the histogram of data points contributing to each altitude. Note that even though the histogram is very skewed toward lower altitudes, the overall agreement with absolute measurements is still present throughout the height profile. Panel (b) shows the full altitude range of the rocket profile versus the limited range of the height profiling technique.
MSIS-90 and found good agreement. This implies that MSIS-90 would also be sufficient for indicating altitudes within this study's region of interest. It is well known that the 558 nm emission comes from a Chapman-like layer whose peak spans many kilometers in altitude. The process of using MSIS to assign a single emission altitude is clearly an oversimplification; ideally, the height would be derived by an appropriately altitude-weighted average of MSIS temperatures matched to the experimental data. This more general altitude weighting has not been implemented in the present algorithm's altitude look-up steps. While such an extension is certainly feasible, it is expected that any resulting correction in altitude would be small compared to other sources of experimental error.

The accuracy of the MSIS model itself has been cross compared with SDI Doppler temperatures in several different ways (e.g., Kaeppler et al., 2015) which have shown strong consistency. Clearly, nightly departures of the temperature profile from its climatological average for the prevailing geophysical indices may occur due, for example, to geophysical activity. However, such deviations should have only small effects on the inferred altitudes. This is because the strong vertical gradient in the climatological temperature profile would require large modeling errors before the resulting altitude errors were significant. For modest values of the Ap and F10.7 indices, at 125 km altitude, there is approximately 5 km of vertical error for every 50K of temperature error in the MSIS model (Figure 2d). The 50K temperature change would correspond to a change in F10.7 of approximately 120, or a change in Ap of approximately 45. These are large differences relative to the uncertainties involved, and differences of this magnitude are unlikely to arise in practice. Even in the case that changes of this magnitude do arise, the effect would largely be to shift the entire altitude range of the height profile; however, the general shape of the profile would remain the same. Overall, even during extended periods of active aurora, the temperature variations in the E-region as a result of magnetic activity are not expected to introduce a significant error (less than 5 km) in altitudes inferred here. Therefore, errors that might arise during the assignment of altitudes are smaller than the altitude resolution of the final product from this technique. Further, the seventh order fitting cannot provide adequate resolution to recover small-scale features.

Currently, the most commonly used empirical model to predict thermospheric neutral winds is the Horizontal Wind Model (HWM). Similar to MSIS, this model has also been through multiple iterations. The most current major version is HWM14. HWM uses data from ground-based and space-based instruments to fit the climatological behavior of neutral winds as a function of latitude, longitude, altitude, day of year, and time of day.

**Figure 10.** (a) Height profile generated from data within the time period used to generate Figure 9a. Plot formats are the same as Figure 9a and replicated here for easy comparison. (b) Height profile generated using a more restricted time period as indicated on the figure.
Figure 12 shows the current HWM model (HWM14) prediction compared to the JOULE rocket mission absolute wind measurements and the optical data. It is clear that the 1D technique agrees with the absolute measurements much better than the HWM model does in this case. At smaller vertical scales, the actual altitude structure of the winds, and particularly wind shears in the lower E-region, are known to be much more complicated than either HWM or the 1D technique can resolve (e.g., Larsen, 2002). Nevertheless, the 1D technique does represent the large-scale behavior reasonably well, and more realistically than HWM, at least for this particular instance.

Figure 11. (a) Results of the monostatic wind fitting technique showing a uniform wind field (blue arrows) as derived from the 558 nm emissions. The cluster of arrows near the center of the field of view suggesting locally stronger winds is likely an artifact due to the small line-of-sight component of a horizontal wind observed in look directions near the zenith. Red to blue hues show varying temperature within the SDI field of view. (b) Example nonuniform wind field using the same fitting technique as (a). Both assume a single emission altitude for all measurements. (c) Harmonic fit of one fundamental Fourier component of the azimuthal variation of the line-of-sight wind observed at 7:26 UT corresponding to the wind field in (a). Fitted winds (red) and observations (white) correlate well. (d) Separate harmonic fitting of one fundamental Fourier component of the azimuthal variation of the wind at 5:10 UT corresponding to the wind field in (b). The Fourier fitting is done independently at each time and represents the best fit possible even if the underlying data might not be a good match. Note the disagreement between the fitted (red) and observed (white) winds in panel (d).

(Drob et al., 2015). Figure 12 shows the current HWM model (HWM14) prediction compared to the JOULE rocket mission absolute wind measurements and the optical data. It is clear that the 1D technique agrees with the absolute measurements much better than the HWM model does in this case. At smaller vertical scales, the actual altitude structure of the winds, and particularly wind shears in the lower E-region, are known to be much more complicated than either HWM or the 1D technique can resolve (e.g., Larsen, 2002). Nevertheless, the 1D technique does represent the large-scale behavior reasonably well, and more realistically than HWM, at least for this particular instance.

An essential component for using this technique is the need for the 558 nm emission to vary in altitude. Time periods when there is not sufficient altitude coverage for the 1D height profiling technique are already amenable to using traditional methods to create reliable 2D maps of the horizontal wind field. The height profiling technique is thus complementary, because time periods that would previously have been rejected can now be used instead to infer 1D height profiles of the horizontal wind. Combined use of the two approaches first makes more complete use of the observational data, and second allows a new capability, that is, to resolve height variations.
Even with the limitations of this technique, it still provides valuable insight by deriving new information from old data in a difficult region to study. More so, current models such as HWM consistently underestimate the variability of winds in this region by a significant amount (e.g., Larsen, 2002; Larsen & Fesen, 2009). By contrast, the optical height profiling technique performed much better in the instance that was able to be validated, as seen in Figure 12. This further emphasizes the need for improved models within this region. The limitations encountered when applying the 1D height profiling technique have motivated new work, currently in progress, to develop an extended fitting technique that will account for variation across longitude, latitude, altitude, and time. This 4D wind mapping technique will thus resolve variations in all geophysical dimensions and will be presented in subsequent papers.

6. Further Application

Examples shown of the 1D height profiling technique have focused on the behavior of E-region winds in the auroral zone. This is a region of complex electrodynamic forcing that produces a range of phenomena and interesting dynamics, such as the shear of horizontal winds in the vertical direction observed by TMA releases (Larsen, 2002). Another region of complex space weather forcing is Earth's geomagnetic cusp. Of all the regions of Earth's thermosphere, the footprints of the geomagnetic cusps are places that are most exposed to direct entry of solar wind plasma and plasma waves. Further, anomalous mass density specific to the upper thermosphere

Figure 12. Comparison of zonal and meridional wind components derived from the JOULE rocket (red) with those obtained from the height profiling technique (black) and those produced by HWM (blue).
in the cusp region has previously been observed and well documented, for example, by H. Liu et al. (2005) and Schlegel et al. (2005). While this particular anomaly does not appear to penetrate down to the E-region, it does demonstrate that the cusp is a region of complex dynamics that may well perturb the lower thermosphere in other ways.

During the early 2000s, satellite accelerometer observations at about 400 km altitude showed a persistent and significant increase in neutral mass density near the thermospheric footprints of the geomagnetic cusps, relative to their larger-scale surroundings. These discovery observations showed the existence of a phenomenon now known as the cusp-region thermospheric mass density enhancement shown, for example, in H. Liu et al. (2005), Figure 5. It is described thoroughly by references previously cited. The data indicate that the mass density within the thermospheric footprint of the cusp is a factor of 1.3–2 greater than in surrounding regions. Lühr et al. (2004) defined the latitudinal width of the anomaly, approximately 350 km, to be the spatial extent of the region across which the mass density exceeds half the maximum density at the peak of the anomaly. Fairly straightforward consideration of the momentum equation suggests that a localized enhancement of this magnitude should be
accompanied by very substantial wind perturbations. Perturbed wind signatures do indeed appear in first-principal models that reproduce the density enhancement, as seen for example, in Figure 3 of Deng et al. (2013). However, we have examined data from Dynamics Explorer 2 (DE-2), shown in Figure 13. We found that no large horizontal wind perturbations associated with the density enhancement were found within the 400–550 km altitude range examined, at least in this climatologically averaged sense. Therefore, the nature of the force balance that sustains the density anomaly remains unclear. While the cusp density enhancement occurs 4–6 scale heights above the altitudes targeted by the optical height profiling technique, the possibility exists that wind anomalies resulting from the forcing associated with cusp processes may nevertheless be observable throughout the thermospheric column.

The lack of wind perturbations seen by DE-2, where models do predict their occurrence, further indicates that the dynamical processes within the cusp are not yet fully explained. The density enhancement is likely to be just one signature of a complex system of space weather perturbations in this region. Using the 1D technique described, along with traditional SDI reconstruction methods, it is possible to examine the 100–150 km altitude range and 240 km altitude for the existence of anomalous wind signatures. This investigation could place a lower bound on where the unique electrodynamics of the cusp cause localized complex behavior with the thermosphere. Previous studies of winds in the cusp over the green-line emission altitude range are limited. The complex and unusual

Figure 14. Height profiles generated from McMurdo (top) and South Pole (bottom) for approximately 1 hr periods centered around times before (left), during (center), and after (right) cusp passage. Despite varying emission altitude, and therefore characteristic energy, the profiles maintained the same general shape throughout the passage of the cusp.
nature of space weather in the cusp has historically made it an area of considerable research. Indeed, the international scientific community deemed the cusp region to be of such importance that in 2018 the “Grand Challenge Initiative: Project CUSP” was introduced as a large international collaborative study of neutral and ionospheric phenomena associated with the cusp regions (Moen et al., 2018).

The thermospheric footprint of the cusp only occurs within a narrow band of geomagnetic latitudes; wind signatures would thus be expected to occur within that same band. Fortunately, there are two SDI sites in the southern hemisphere that are located within these latitudes: South Pole and McMurdo Stations in Antarctica (Table 1). The cusp footprint typically passes over these instrument locations for up to a few hours each night. A number of days were examined from the 2016 and 2017 observing seasons. Figure 14 presents one representative example, showing comparisons from before, during, and after the cusp region passed over the two observational sites. It shows that no significant signatures were introduced on this day into the wind’s altitude profiles during times when the cusp would be expected to be over the observation sites. Further, we did not find any days for which similar analysis yielded conspicuous signatures of cusp-related wind anomalies in the E-region.

Conclusions from results shown in Figure 14 are limited to the height ranges available to model with this technique and thus only range up to ~150 km. However, the 630 nm oxygen emission can be used to test whether any perturbations are observed at a higher altitude, specifically 240 km. By generating time-evolving wind field plots for 630 nm oxygen emissions, any perturbation at these altitudes should be detected. Figure 15 shows a detailed example of the 630 nm oxygen emission wind field plots. We examined data from the Antarctic winter observing seasons of 2016 and 2017 and, once again, no obvious wind perturbations were apparent during times of cusp passage.

Despite the lack of any anomalies found in neutral winds at altitudes of 240 km and below, these results still constrain the altitude range over which such signatures could be present. The lack of wind anomalies detected in this study suggests that wind perturbations associated with the cusp will most likely be found between 240 and 400 km altitude. Below 240 km, satellites have very limited orbital lifetimes. However, this study strongly indicates that any wind anomalies that do exist must occur at altitudes above those studied here. These altitudes are densely populated by low Earth-orbiting satellites, making a complete understanding of the relevant mechanisms more important as the associated processes may limit the accuracy of orbital predictions. In particular, high accuracy is required for applications such as collision avoidance and re-entry calculations.

7. Conclusions

This study shows a fundamentally new capability in which optical Doppler remote sensing is used to derive height profiles of horizontal wind in Earth’s E-region under appropriate conditions. This method provides the ability to use 558 nm Doppler spectra from times that might otherwise be rejected because of altitude variations with time and with look direction. Alternatively, periods of time without altitude variation in the emissions can now be reconstructed reliably using traditional wind fitting techniques. However, under these circumstances, there is little information available about the altitude variations of the wind.

Results from the 1D height profiling technique were compared to absolute in-situ wind data from the JOULE rocket mission. On this night, altitude variations were not ideal for the 1D height profiling technique presented...
in this paper. However, results agreed well with data from the JOULE rocket which is a known reliable indicator of the winds. In many of the time periods examined during this study, the sampling at upper altitudes appeared sparse relative to the observational coverage at lower altitudes. Practical experience showed that even in cases where the upper heights appeared sparsely illuminated by the aurora, they typically still included tens of data points per height bin. Further, these tended to be spread widely across the field of view, thus ensuring that the polynomial height profile fitting remained well-conditioned. Stability during less-than-ideal time periods was validated by choosing a number of subsets of the line-of-sight wind data, with each being used to generate a separate height profile. All such profiles were found to be consistent with the JOULE data. While agreement was found between vertical profiles of the horizontal wind derived by the optical and rocket-based techniques, they were not consistent with height profiles from HWM.

One limitation is that this technique cannot match the height resolution of the rocket profiles. Rather, its major application is to define the overall shape of the vertical profile of the zonal and meridional winds. Another potential concern is associated with possible inaccuracy of the MSIS model that is used to assign an altitude to each observation. Experience from numerical testing shows that artifacts caused by nightly deviations from the MSIS climatological average temperature profile do not in practice have a large impact on final fitted wind profiles. Lastly, monostatic wind fits must be used in advance, to test whether nonuniform horizontal winds are likely to be present and thus cause artifacts in the 1D technique. Despite the caveats, and even after excluding time periods of horizontal nonuniformity, there are many instances when the technique can provide wind...
profiles at less expense and with higher cadence than is possible with the only other available method, that is, rocket-borne chemical releases. Further, this technique has established the necessary background information for more complex wind mapping algorithms currently under development. In particular, it confirmed the possibility of implementing a 4D wind mapping technique that will be presented in subsequent papers.

A practical application of the height profiling technique was shown by applying it to periods around cusp passage over instruments located in Antarctica. The cusp region is host to complex space weather forcing that is not well understood. Further, anomalous behavior has been previously observed in the cusp. The technique was used to search for any lower thermospheric signatures of anomalous winds associated with the footprint of the cusp. Results showed no detectable anomalies during the cusp passage at the E-region altitudes that are the focus of this study, nor at F-region altitudes where the 630 nm emission originates. These results are broadly consistent with previous results from DE-2. Further, the results place a lower limit on the altitude region at which the complicated dynamics of the cusp footprint may impact neutral flows.

Data Availability Statement

DE-2 Unified Abstract data were provided courtesy of NASA Space Science Data Center Coordinated Archive (https://nssdc.gsfc.nasa.gov/nmc/dataset/display.action?id=SPIO-00040). SuperDARN measurements of ion convection velocities shown in Figure 16 were provided by the SuperDARN collaborative courtesy of W. A. Bristow (personal communication 2016). SSUSI data shown in Figure 16 were provided by the Johns Hopkins University Applied Physics Laboratory courtesy of L. J. Paxton (https://ssusi.jhuapl.edu/data_availability). SDI data are available at http://sdi_server.gi.alaska.edu/sdi_web_plots/sdi_arc.asp.

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