Lower thermospheric wind estimates from dual-beam incoherent scatter radar measurements

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Abstract. A method for estimating the vector neutral wind profiles in the mesosphere and lower thermosphere (MLT) region of the upper atmosphere from Arecibo dual-beam incoherent scatter radar data is presented. The method yields continuous estimates of both the altitude-averaged *F*-region plasma drifts and the altitude-resolved neutral wind profiles in the MLT using data taken while the Arecibo feed system swings in azimuth. The problem is mixed determined, and its solution is not inherently unique. Second order Tihkonov regularization is used to find solutions consistent with the available data while being minimally structured, additional structure being unsupported by the data. The solution is found using the method of conjugate gradient least squares and sparse matrix mathematics. Example data acquired during an interval of midlatitude spread *F* are used to illustrate the method.

1. Introduction

Among the most important state parameters for the study of the mesosphere and lower thermosphere (MLT) region are the neutral wind velocities. The winds are indicators of large-scale circulation in the upper atmosphere, of energy and momentum being transported across it by waves, and of the overall dynamical stability. Winds can be measured accurately in the MLT using chemical releases and by high-power Doppler lidar (e.g. *Larsen* [2002]; *Zhao et al.* [2003]; *Chu et al.* [2005]). The relative ease of accessibility and geographic diversity of the world's incoherent scatter radars (ISRs), deployed at low, middle, and high latitudes on several continents, makes them attractive instruments for MLT wind observations as well. However, complications arise since it is the ionized rather than the neutral gas motion that the radars observe directly, and a number of different techniques appropriate for different altitude regimes have been applied to the neutral wind problem. The purpose of this paper is to describe a new method for estimating vector neutral winds in the MLT using the Arecibo Radio Telescope, the most sensitive of the ISRs. As Arecibo can only observe drifts in two coplanar directions at a time, statistical inference is a necessary component of the technique.

In fact, Arecibo has been making wind measurements in the MLT for quite a few years. Most of the important principles involved are discussed by *Harper* [1977] and references therein. In the ionosphere, the neutral wind in the direction parallel to the geomagnetic field is a calculable function of the parallel-to-B ion drift speed. Below about 170 km, the two speeds are essentially equivalent, while above this, ion diffusion must also be considered and accounted for [*Vasseur*, 1969; *Behnke and Kohl*, 1974]. By measuring line-of-sight ion drifts with two beams in the magnetic meridional plane and assuming horizontal invariance, it is possible to isolate the parallel-to-B ion drift component and to estimate the associated neutral wind component. With the assumption that the vertical neutral wind is small, the meridional wind profile can thus be calculated.

The relationship between ion drifts perpendicular to B and the winds is more complicated, and winds can only be estimated from incoherent scatter below about 130 km where the ions are unmagnetized [*Harper et al.*, 1976]. An estimate of the perpendicular-to-B electric field is required to separate the Pedersen and Hall drift components from the measured ion drifts and to isolate the contribution due to neutral winds. The required E-field estimate can be formulated on the basis of ion drifts measured in the fully-magnetized *F* region along three or more non-coplanar bearings, assuming spatial homogeneity and efficient electric field mapping along magnetic field lines (e.g. *Behnke and Harper* [1973].)

Estimating neutral winds in the perpendicular-to-B direction also requires a specification of the the composition and the ion-neutral collision frequency for each important ion species. In practice, isomorphism in the incoherent scatter spectrum makes it impractical to estimate temperature, composition, and collisionality by fitting range-gated ISR data, even below 130 km altitude where the electron and ion temperatures are approximately equal. Modeling is generally incorporated in this aspect of the methodology. Below about 110 km altitude, however, the collisional coupling between ions and neutrals is strong, and ion and neutral drift velocities can be regarded as being essentially equivalent.

Variations of one aspect or another of the Harper methodology have been employed by numerous investigators for wind and wave studies in the midlatitude MLT region (e.g. *Burnside et al.* [1983]; *Roper et al.* [1993]; *Aponte et al.* [2005]; *Gong et al.* [2012]; *Huang et al.* [2012]). These include studies that focus on the mesosphere, where the procedure is especially straightforward [*Mathews et al.*, 1981; *Rottger et al.*, 1981; *Fukao et al.*, 1982; *Rottger et al.*, 1983; *Fukao et al.*, 1985; *Tsuda et al.*, 1985; *Maekawa et al.*, 1986; *Zhou and Morton*, 2006]. The techniques and related databases have become sufficiently reliable to support studies of long-term trends over Arecibo [*Maekawa et al.*, 1987; *Tepley et al.*, 2011; *Santos et al.*, 2011; *Brum et al.*, 2012].

Harper [1977] utilized a multipulse pattern to make range-resolved autocorrelation function measurements in the *E* region and a long-pulse pattern to make accurate line-of-sight velocity measurements in the *F* region. Contemporary experiments at Arecibo have replaced the multipulse mode with a coded long pulse (CLP) and the long pulse with a multifrequency mode (MRACF) — see *Sulzer* [1986a, b] for details. The new pulsing schemes exploit the capabilities of the Arecibo Observatory more fully than the old ones.

Wind measurements at Arecibo have generally been based on ISR data collected with either a single beam or with two coplanar beams directed at fixed azimuths. The fixed-azimuth approach is somewhat inefficient, however, since time spent during beam swinging maneuvers is wasted. Moreover, there is an implicit assumption that the winds and electric fields in the upper atmosphere are invariant during the time that it takes to acquire data and swing the beam. In view of the slow slew rate of the Arecibo beam system, this assumption may not be difficult to violate. *Sulzer et al.*

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[2005] introduced a novel refinement, incorporating statistical inverse theory in the estimation of *F*-region plasma drifts from data taken when the beam was continuously swinging. In this paper, we generalize the result and apply a related technique to the entire horizontal wind estimation problem. The current work is an expansion of the more basic technique reported on by *Hysell et al.* [2009]. It bears some similarities to the work of *Nygrén et al.* [2011] except that we treat the under-determined problem using a damped least-squares approach whereas they treated the even- or over-determined problem using least squares.

2. Methodology

Here, we describe a method to infer the vector neutral winds in the MLT region from Arecibo dual-beam incoherent scatter data. Ionospheric electric fields are estimated on the basis of *F*-region incoherent scatter ion-line line-of-sight drift measurements made using the MRACF mode [*Sulzer*, 1986a], which are averaged over all *F*-region altitudes. Wind estimates are derived using *E*-region incoherent scatter ion-line line-of-sight drift measurements made using the coded long-pulse (CLP) mode [*Sulzer*, 1986b], which are range resolved. (The range resolution of the CLP data used here is 300 m.) The aforementioned electric field estimates are incorporated into the procedure for estimating winds. Information about ion composition is also required and is extracted from CLP autocorrelation function estimates through an iterative fitting procedure that involves a simple modeling component. The cadence of the experimental results is approximately once per minute.

Throughout the subsequent discussion, we make use of the language and formalism of statistical inverse theory as presented by *Menke* [1984]; *Tarantola* [1987]; *Aster et al.* [2005].

2.1. Electric field estimates

Ion drift measurements in the *F* region ionosphere contain information about the transverse electric field. We regard this field to be essentially invariant throughout the *F* region and combine measurements from different range gates to improve statistical accuracy. Extracting electric field information entails performing some coordinate transformations and then implementing an inverse method.

The line-of-sight *F* region drifts measurements acquired with the linefeed and Gregorian beam systems at Arecibo can be related to drifts in a geographic Cartesian (up-east-north) coordinate system with the assumption of spatial homogeneity through the linear transformation

$$\begin{pmatrix} v_l \\ v_g \end{pmatrix} = \underbrace{\begin{pmatrix} 1 & & \\ c\theta & s\phi s\theta & c\phi s\theta \end{pmatrix}}_{Q} \begin{pmatrix} v_u \\ v_e \\ v_n \end{pmatrix} \tag{1}$$

where θ and ϕ are the azimuth and zenith angle of the Gregorian feed. (The linefeed is taken to point toward zenith, although a more generalized form of (1) could be formulated trivially.) Of all the angles used in this analysis, only θ varies in time in practice. We use "s" and "c" as shorthand for sin and cosine.

More useful for the analysis than the geographic coordinate system is a Cartesian system aligned with the geomagnetic field. We define a right-handed system with a component parallel to $\mathbf{B}$, a component perpendicular to $\mathbf{B}$ in the horizontal plane, and another perpendicular component in the plane of the magnetic meridian. The components of the velocity in this system ($v_\parallel$, $v_{\perp e}$, $v_{\perp n}$) are related to the components in geographic coordinates through the transformation:

$$\begin{pmatrix} v_\parallel \\ v_{\perp e} \\ v_{\perp n} \end{pmatrix} = \underbrace{\begin{pmatrix} c\xi & s\eta s\xi & c\eta s\xi \\ & c\eta & -s\eta \\ s\xi & -s\eta c\xi & -c\eta c\xi \end{pmatrix}}_{R^{-1}} \begin{pmatrix} v_u \\ v_e \\ v_n \end{pmatrix} \tag{2}$$

where η and ξ are the zenith and azimuth angle of the magnetic field, respectively. The transformation above can be inverted, yielding

$$\begin{pmatrix} v_u \\ v_e \\ v_n \end{pmatrix} = \underbrace{\begin{pmatrix} c\xi & & s\xi \\ s\eta s\xi & c\eta & -s\eta c\xi \\ c\eta s\xi & -s\eta & -c\eta c\xi \end{pmatrix}}_{R} \begin{pmatrix} v_\parallel \\ v_{\perp e} \\ v_{\perp n} \end{pmatrix} \tag{3}$$

All together, the measured line-of-sight drifts can then be related to the drifts in magnetic coordinates with the definition of the transformation matrix $S \in \mathbb{R}^{2\times 3} = QR$. The nonzero elements of S are given by:

$$\begin{aligned} S_{11} &= c\xi \\ S_{13} &= s\xi \\ S_{21} &= c\theta c\xi + s\theta s\xi(s\phi s\eta + c\phi c\eta) \\ S_{22} &= s\theta(s\phi c\eta - c\phi s\eta) \\ S_{23} &= c\theta s\xi - s\theta c\xi(s\phi s\eta + c\phi c\eta) \end{aligned}$$

such that

$$\begin{pmatrix} v_l \\ v_g \end{pmatrix} = S \begin{pmatrix} v_\parallel \\ v_{\perp e} \\ v_{\perp n} \end{pmatrix}$$

The preceding analysis accommodated a single pair of line-of-sight drift measurements acquired at a single time. For n measurements acquired at n distinct times, the resulting $2n$ data are related to the corresponding $3n$ drift components by:

$$\begin{pmatrix} v_{l1} \\ v_{g1} \\ \vdots \\ v_{ln} \\ v_{gn} \end{pmatrix} = \underbrace{\begin{pmatrix} S_1 & & \\ & \ddots & \\ & & S_n \end{pmatrix}}_{T\in\mathbb{R}^{2n\times 3n}} \begin{pmatrix} v_{\parallel 1} \\ v_{\perp e1} \\ v_{\perp n1} \\ \vdots \\ v_{\parallel n} \\ v_{\perp en} \\ v_{\perp nn} \end{pmatrix} \tag{4}$$

where the subscripts refer to time steps and all the drifts measurements are taken to be independent.

The transformation in (4) has the form of the standard forward problem $d = Gm$, d being the data, the dual-beam drift velocity measurements at n different times in this case, m being the model parameters or state, the underlying vector drifts in a magnetic coordinate system, and "G" being the calculable, time-dependent linear system that transforms from one to the other. The matrix T is not square, the inverse problem is underdetermined, and the model parameters m cannot be found uniquely on the basis of the given data d as written. In order to proceed, we pose the inverse problem as an optimization problem, where additional penalties are introduced to reduce the solution space of m, creating a unique solution.

Applying the method of weighted, damped least squares, we seek a solution for m that minimizes the weighted forward prediction error with an added damping factor representing the additional penalty:

$$m = \operatorname*{argmin}_m \, (Gm - d)^t W_d (Gm - d) + \alpha^2 m^t W_m m \tag{5}$$

where W_d is the inverse data error covariance matrix and W_m is a model weight matrix. If W_m is made to be the identity matrix, then the damping term plays the role of minimizing the model length, which can be interpreted as an implementation of Occam's Razor. Other choices for W_m can be used to minimize the global first or second derivative (the roughness) of the model parameters. These choices are tantamount to performing zeroth-, first-, or second-order Tihkonov regularization. The α term is a regularization parameter which balances the impacts of the penalties.

Using the method of normal equations, it is easy to show that the inverse operator that satisfies the optimization equation is given by the weighted-damped least squares operator

$$\hat{G}^{-1} = (G^t W_d G + \alpha^2 W_m)^{-1} G^t W_d \quad (6)$$
$$m \approx \hat{G}^{-1} d \quad (7)$$

Applying the operator above to the data yields estimates of the vector ion drifts in a magnetic coordinate system at different times. The transverse components of the drifts are indicative of $\mathbf{E} \times \mathbf{B}$ drifts. The parallel component is not of interest to the present analysis and can be discarded.

2.2. Neutral wind profile estimates

In the E region ionosphere/ lower thermosphere, ion drift measurements are indicative of electric fields and neutral winds. As with the electric fields, estimating the neutral winds from the measurements involves a series of coordinate transformations followed by an inverse method. It also involves additional algebraic manipulations as well as a correction for (really the removal of) the electric field signature. Finally, we do not regard the vector neutral winds as being invariant with altitude and instead seek to estimate neutral wind profiles from the data.

The ion drifts in the E region are related to the electric fields and winds by the following formulas:

$$\begin{aligned}
v_\parallel &= u_\parallel \\
\mathbf{v}_\perp &= \frac{\mathbf{E}\times\mathbf{B}}{B^2} r_\circ + \frac{\mathbf{E}_\perp}{B} r_1 + \mathbf{u}\times\hat{b} r_1 + \mathbf{u}_\perp r_2 \\
r_\circ &= \frac{1}{1+\nu^2/\Omega^2} \\
r_1 &= \frac{\nu/\Omega}{1+\nu^2/\Omega^2} \\
r_2 &= \frac{\nu^2/\Omega^2}{1+\nu^2/\Omega^2}
\end{aligned}$$

where ν is the ion-neutral collision frequency and Ω is the ion gyrofrequency. Both of these parameters are functions of altitude to be modeled. For the aforementioned F-region analysis, $1 \sim r_\circ \gg r_1 \gg r_2$, but for the E-region, the correct ordering is $1 > r_\circ \sim r_1 \sim r_2$. In the event of multiple ion species, the correct prescription is to calculate the various r_i for each species separately and then to add the results, weighted by the fractional number density of the given species. Ion mobilities add fractionally. Below, the determination of the ion composition is considered as a separate problem.

We regard the ionospheric electric fields as being spatially invariant and use our F-region results as the basis for their estimation. In matrix notation, the E-region ion drifts can be related to the F-region drifts and the neutral winds in the lower thermosphere as follows:

$$\begin{pmatrix} v_\parallel \\ v_{\perp e} \\ v_{\perp n} \end{pmatrix}_E - \underbrace{\begin{pmatrix} v_{\perp e} r_\circ + v_{\perp n} r_1 \\ v_{\perp n} r_\circ - v_{\perp e} r_1 \end{pmatrix}_F}_{U}$$
$$= \underbrace{\begin{pmatrix} 1 & & \\ & r_2 & -r_1 \\ & r_1 & r_2 \end{pmatrix}}_{V} \begin{pmatrix} u_\parallel \\ u_{\perp e} \\ u_{\perp n} \end{pmatrix} \quad (8)$$

Here, the E and F subscripts indicate E- and F-region ion drifts. Multiplying both sides of (7) gives a relationship between the data and the underlying state parameters:

$$\begin{pmatrix} v_l \\ v_g \end{pmatrix}_E - SU = SV \begin{pmatrix} u_\parallel \\ u_{\perp e} \\ u_{\perp n} \end{pmatrix} \quad (9)$$
$$\begin{pmatrix} v'_l \\ v'_g \end{pmatrix} = Y \begin{pmatrix} u_\parallel \\ u_{\perp e} \\ u_{\perp n} \end{pmatrix} \quad (10)$$

where we have defined $Y \in \mathbb{R}^{2\times 3} = SV$. In our notation, primed quantities are line-of-sight drift measurements in the lower thermosphere with corrections for electric fields having been made.

Whereas (9) applies for a measurement at a single time and range, we can write the result more generally for the case of data acquired at n times and m ranges independently:

$$\begin{pmatrix} v'_{l1,1} \\ v'_{g1,1} \\ \vdots \\ v'_{lm,n} \\ v'_{gm,n} \end{pmatrix} = \underbrace{\begin{pmatrix} Y_{1,1} & & \\ & \ddots & \\ & & Y_{m,n} \end{pmatrix}}_{Z\in\mathbb{R}^{2mn\times 3mn}} \begin{pmatrix} u_{\parallel 1,1} \\ u_{\perp e1,1} \\ u_{\perp n1,1} \\ \vdots \\ u_{\parallel m,n} \\ u_{\perp em,n} \\ u_{\perp nm,n} \end{pmatrix} \quad (11)$$

which again has the form of the equation $d = Gm$. While the problem could be inverted with the weighted-damped least squares approach, the size of the matrix X is now prohibitive. Instead, we will employ an iterative method, the method of conjugate gradients least squares or CGLS, making use of the tools of sparse mathematics to improve computational performance. One of the benefits of CGLS is that it does not require any matrix inverses or even matrix-matrix multiplies within the iterative loop.

The method of normal equations again shows that the solution to (5) is given given by $(G^t W_d G + \alpha^2 W_m) m = G^t W_d d$ or canonically as

$$\begin{aligned}
\tilde{G}^t \tilde{G} m &= \tilde{G}^t d \quad (12) \\
\tilde{G} &\equiv \begin{pmatrix} W_d^{1/2} G \\ \alpha W_m^{1/2} \end{pmatrix} \\
\tilde{d} &\equiv \begin{pmatrix} W_d^{1/2} d \\ 0 \end{pmatrix}
\end{aligned}$$

where the fact that the weight matrices are positive definite has been utilized in their factorizations. This solution has the form $Ax = B$ required for the application of the CGLS method. Note crucially that the method is applied to the canonical equation (12) and never to the equation $\tilde{G}m = \tilde{d}$, which only holds approximately and which cannot be solved stably other than by optimization methods such as the one outlined here.

Finally, we can use R to express the vector wind estimates in geographic coordinates.

2.3. Ion composition

It is well known that least-squares fitting of the incoherent scatter spectrum cannot be used to distinguish ion composition in the lower thermosphere because of unfortunate isomorphism in the theoretical shape of the spectrum. Different combinations of atomic and molecular ions at different temperatures yield essentially identical incoherent scatter spectra. Ion neutral collisions make the uniqueness problem more severe. One approach to the problem is to use a standalone, empirical model for the temperature or the composition and to fit for the remaining parameters. Lacking a suitable model, we adopt another approach that utilizes a simple theoretical model for the composition. In the altitude region of interest, we may take $T_e = T_i$, simplifying the problem considerably.

In the lower thermosphere, we neglect transport and assume that all of the ions are in photochemical equilibrium. We consider only O^+, O_2^+, and NO^+ ions, the latter two being produced

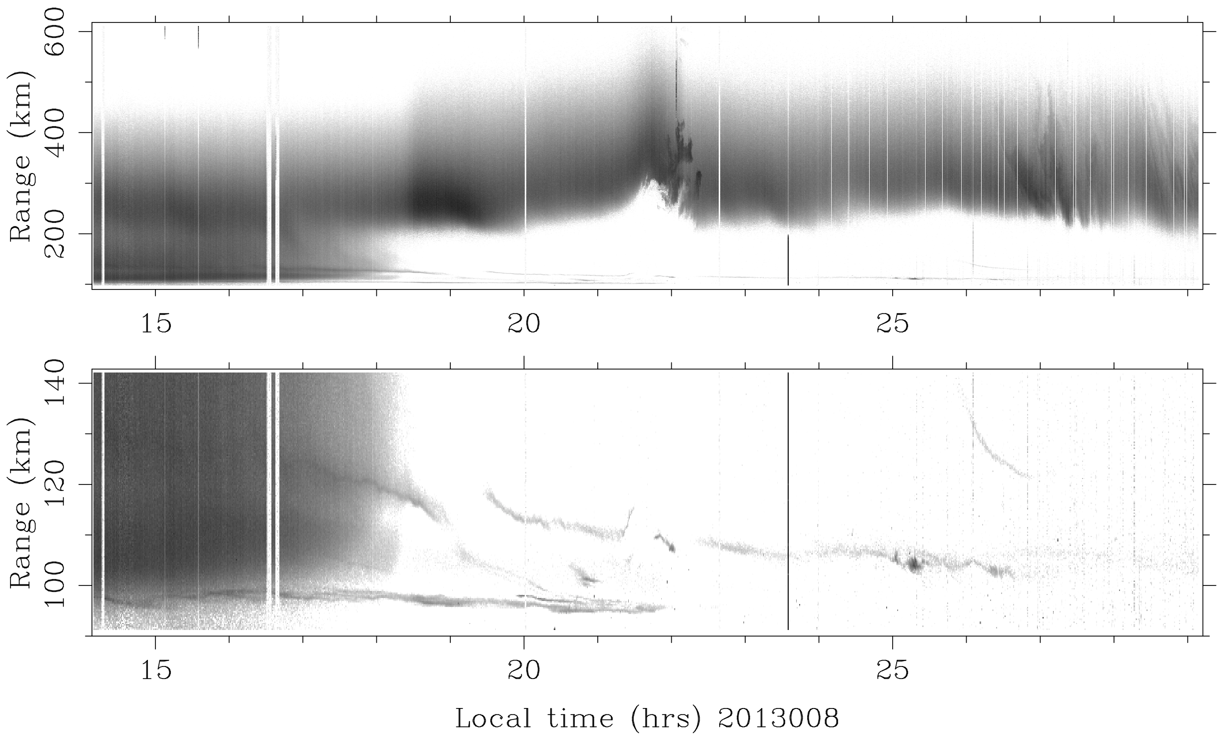

Figure 1. Range time intensity (RTI) plot of Arecibo line feed data for the evening of January 8, 2013. The received signal-to-noise ratio, scaled by a factor of the range squared, is plotted as a proxy for electron density on an arbitrary scale. The upper panel shows echoes from all ranges,and the lower panel highlights echoes from the *E* region.

through charge exchange reactions involving the former and destroyed through dissociative recombination. Consequently, we may write the following equations balancing production (lhs) and loss (rhs):

$$\alpha N[O^+]N[O_2] = \gamma N[O_2^+]N[e^-] \quad (13)$$

$$\beta N[O^+]N[N_2] = \delta N[NO^+]N[e^-] \quad (14)$$

where α and β (γ and δ) are the rate coefficients for charge exchange (dissociative recombination). Up-to-date expressions for these coefficients can be found in *Schunk and Nagy* [2004]. Neutral density estimates are taken from the NRLMSISE-00 model *Picone et al.* [2002].

Equation (13) and (14) can be combined and solved algebraically for the number density fraction of the molecular ion species and, by inference, of O^+. These fractions are used in the computation of the ion mobilities needed in section 2.2. Ion-neutral collision frequencies are likewise estimated with the help of theoretical formulas taken from *Schunk and Nagy* [2004]. Since the coefficients of dissociative recombination depend on the electron temperature, this parameter must be estimated from the incoherent scatter spectra, which are fit using an iterative process that utilizes the composition and collision frequency information. The electron density, meanwhile, is estimated on the basis of the incoherent scatter power, normalized to a constant determined from ionograms recorded nearby. Since the altitudes of interest are in Arecibo's near field, we take the backscatter power to be proportional to the electron density and to be invariant with range.

3. Example data

The event in question is presented in Figure 1, a range-time-intensity (RTI) representation of the incoherent scatter signal received with the zenith-looking line-feed system at Arecibo on the evening of Jan. 8, 2013. The figure shows the echo signal-to-noise ratio, scaled by a factor of the square of the range and intended to serve as a proxy for electron density. The entire sample raster appears in the top half of the figure, whereas the *E*-region samples are shown on an expanded range scale on the bottom half.

The data were collected under geomagnetically active conditions and are indicative of a strong midlatitude spread-*F* event. A plume is visible between 2000–2300 LT, followed by what appear to be medium-scale traveling ionospheric disturbances (MSTIDs) after about 0230 LT the following morning. The *F*-region plumes are reminiscent of what is observed at low magnetic latitudes under equatorial spread *F* (ESF) conditions. An important difference is that the plume here exhibits both locally enhanced and depleted regions of plasma (ESF is characterized predominantly by depletions.)

The *E* region echoes show the presence of two sporadic *E* layers. The first descended from about 120 km at sunset ($\sim$1800 LT) and persisted for approximately nine hours thereafter. The second maintained an altitude below 100 km and terminated by 2200 LT. Neither layer was particularly dense or particularly structured. The higher-altitude layer actually broke up around the time of the spread *F* plume and only exhibited vertical structuring and billowing in the interval between the plume and the appearance of the MSTIDs. No important coupling between the *E* and *F* layers appears to have played a role in this remarkable event.

Figure 2 shows the results of processing the signals depicted in the previous figure. Panels (a) and (c) show *E*-region line-of-sight ion drifts in the *E* region computed from CLP data from the line and Gregorian feeds, respectively. Panels (b) and (d) show range-integrated *F* region line-of-sight ion drifts computed from MRACF data from the line and Gregorian feeds. A dashed line that runs along the bottom of panel (c) shows the azimuth (from 0–360°) of the Gregorian feed pointing. The passage of the midlatitude spread *F* plume around 2200 LT is clearly evident in the MRACF data. Structuring in the Gregorian MRACF drifts signal is evidence of the large contribution of horizontal motion to the line-of-sight drifts observed obliquely.

Close examination of panel (a) shows echoes with small Doppler shifts below about 115 km, red shifts between about 115–125 km, and blue shifts above that altitude. We can interpret these three zones as zones where the ion drifts are dominated by neutral winds, Pedersen drifts, and Hall drifts, in ascending order. This implies that the electric field should have northward and westward perpendicular-to-B components, or westward and southward perpendicular-to-B drift, before 1800 LT.

Even closer examination of panel (c) around 100 km altitude prior to 1800 LT reveals strong variations in the line-of-sight drifts with altitude and azimuth. The entire data record and this area in particular is indicative of vertical shear in the horizontal ion drift velocity, itself indicative of strong wind shear.

Panel (e) shows the temperatures, fits to the line feed CLP autocorrelation functions performed iteratively with the composition model from section 2.3. While the isotherms are approximately horizontal in the panel, significant perturbations are evident, and some of the mixing suggested in the RTI plots in Figure 1 is also suggested by the temperatures. Note that as the sporadic *E* layers are likely metallic in composition, we do not expect our fitting procedure to yield accurate temperature estimates within them or the strata from which they emerged.

The results of the inversion of the MRACF drifts described in section 2.1 are plotted in panel (f). The model prediction error associated with these estimates is small, i.e., the line-of-sight drifts predicted by these estimates resemble the data closely. At the same time, regularization presents severe artifacts in these estimates associated with beam swinging that would otherwise occur. As expected, the estimates suggest westward and southward perpendicular drifts before 1800 LT accompanied with positive parallel (down the field line) drifts that combine to give rise to the descent of the *F* layer. Descent continues until about 2100 LT, when the perpendicular drifts reverse to northward and overcome the downward parallel motion. Just before 2200 LT, the perpendicular drifts turn southward again, and the parallel drifts intensify, leading to the rapid descent and collapse of the *F* layer. The midnight collapse is a regular and well-documented feature of the midlatitude ionosphere [*Gong et al.*, 2012]. In this case, it was unusually dramatic and accompanied by convective plasma instability. Not only the *F* layer but also the upper sporadic *E* layer was affected.

Between 1800–2400 LT and again after 0200 LT, the eastward perpendicular drifts were large. It is noteworthy that the period of strong westward drifts after 0200 LT was coincident with the appearance of MSTIDs.

Finally, panels (g) and (h) show horizontal wind estimates for the period between 1400–1800 when the *E* region was still dense. The estimates were derived using the procedure outlined in section 2.2. Plotter symbols ("X") denote regions of poor data quality, which gave rise to somewhat spurious results.

As anticipated, the horizontal winds exhibit large amplitudes, quasi-periodic variations with altitude, and strong shears in the altitude region of interest. The winds are characterized by vertical wavelengths of about 20 km and slow downward propagation. This is true of both the zonal and the meridional component. The largest amplitudes and strongest shears occur at the bottom of the *E* layer, around 100 km, which is consistent with the characteristics broadly evident in panel (c). Less evident in panel (c) but present nonetheless is a region of strong shear higher in altitude, around 125 km.It is noteworthy that from these two altitude regions emerged the sporadic *E* layers described earlier.

4. Significance and future work

The objective of this work is to optimize the utilization of the Arecibo Radio Telescope, the most sensitive of the incoherent scatter radars, for wind profile measurements in the MLT region. This has been done by replacing multiple fixed-azimuth data with data collected while the azimuth is varying continuously. The problem is mixed determined, and statistical inverse methods are required to overcome problems associated with the existence, uniqueness, and stability of wind profile solutions. The problem is linear and has been addressed here using 2nd-order Tikhonov regularization, a conjugate-gradient least-squares solver, and sparse matrix math.

The slew rate for the Arecibo feed system is 25°/min. or 90° in 3.6 min. A fixed-azimuth experiment designed to spend more time acquiring data in either cardinal direction than it does in beam swinging therefore has a practical cadence of no more than one solution every 10.8 min. However, this interval is comparable to or longer than the Brunt-Vaisala period in the MLT region. Much of the fine structure in the wind profiles is consequently excluded by fixed-azimuth experiments. The analysis described in this paper produced wind profiles every minute, but that figure is essentially arbitrary and is set by the incoherent integration time of the spectral data. The effective cadence of the experiment is limited by the amount of regularization used, a parameter adjusted in this case to limit the variance in the profiles (in altitude and time) without introducing significant model projection error. The evidence of Figure 2 is that the method permits variations in the wind profiles on timescales of 5 min. or less.

Finely-resolved neutral wind measurements in the MLT region are required in view of the predominant wind characteristics that are fundamentally non-tidal (e.g. [*Harper*, 1977; *Larsen*, 2002; *Sherman and She*, 2006; *Liu*, 2007]). The region is the site of wave breaking and wave-mean flow interactions as well as a number of viable instabilities. This is a frontier research area in need of incisive experimental tools.

A number of improvements to the basic methodology described here are possible. A simple one would be to limit the azimuth scans to a single quadrant. This would reduce by a factor of 4 the size of the volume being probed by the radar, making the assumption of spatial homogeneity more likely to be satisfied.

Another improvement involves combining *F*-region plasma drift and *E*-region neutral wind estimation into a single optimization problem. We expect the vertical neutral winds to be relatively small, and this imposes additional constraints on the electric fields, as even small errors in the latter give rise to large deviations in the former. More accurate electric field estimates, in turn, translate to more accurate horizontal wind estimates. The combination of the two inverse problems into one, combined with the addition of a small penalty on the size of the vertical wind estimates, constitute an important potential refinement.

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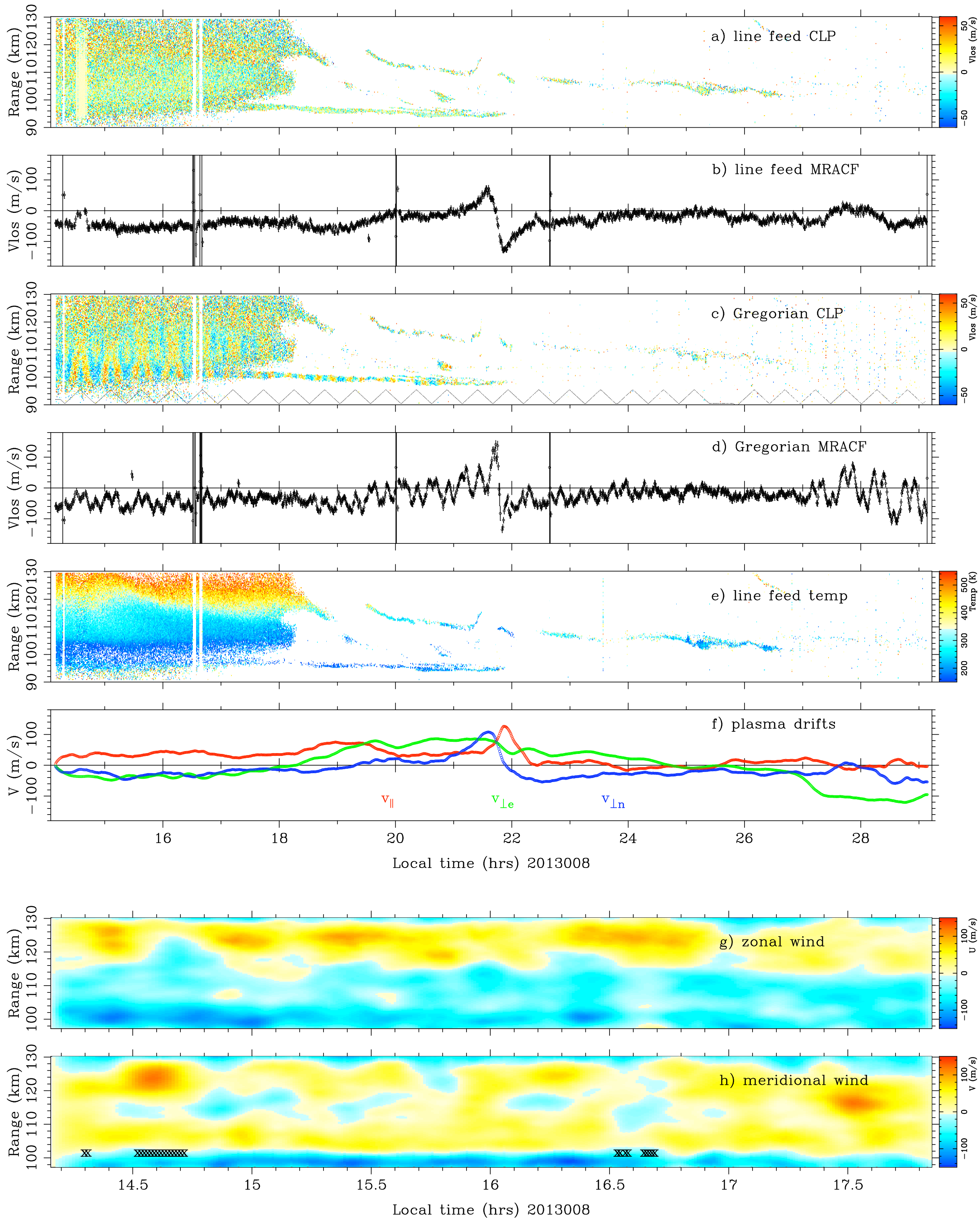

Figure 2. Results from the Arecibo winds analysis.Panels (a) and (c) show line-of-sight ion drifts measured with the coded long pulse technique using the line and Gregorian feeds, respectively. Panels (b) and (d) show averaged F-region line-of-sight plasma drifts measured by the line and Gregorian feeds, respectively. Panels (e) shows temperature estimates from the line feed data. Panel (f) shows derived F-region plasma drifts. Panels (g) and (h) show derived zonal and meridional wind speed profiles, respectively. Note that range is equal to altitude for line-feed data only.

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