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A comparison study of the auroral lower thermospheric neutral winds derived by the EISCAT UHF radar and the Tromsø MF radar.

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Abstract

A comparative study of the neutral wind in the polar upper mesosphere/lower thermosphere is conducted using two radars, the European Incoherent Scatter (EISCAT) UHF radar (ν=931 MHz) and the Tromsø MF radar (ν=2.8 MHz) co-located in northern Scandinavia. Comparisons of winds are made in case studies as well as on a statistical basis. Based on simultaneous observations obtained by the two radars, comparisons over a height range from ~90 to ~100 km are made for about 20 days between February and October, 1999. In these comparisons, we directly compare temporal wind variations. On the other hand, mean winds as well as diurnal and semidiurnal components are derived using 1-month-averaged wind data obtained by the MF radar for 1999. They are compared with those derived from 56 days of EISCAT data studied by *Nozawa and Brekke* [1999a,b]. In the case studies, we have also utilized altitude profiles of electron density obtained by the EISCAT radar at and above 62 km height to determine the total reflection height as well as estimating the effect of group retardation for 2.8 MHz radio wave. It can be seen that the effect of particle precipitation sometimes penetrates into the D region (down to ~84 km). Generally, wind velocities derived by EISCAT with the monostatic method exhibit significant scatter with time below 100 km, while the number of the MF radar wind data sharply decreases above 90 km except for summer. These facts make the comparison difficult, but no significant departures between different radar winds are identified on an instantaneous basis. In the statistical studies, generally both altitude profiles are well connected, or continuous, but significant disagreements are observed for the mean wind and the tidal components for summer. Based on these comparisons, we raise strong concerns regarding the use of summer wind data above 91 km obtained by MF radar located at high latitude under high solar activity conditions. The wind values during the winter and equinox observed by the 2 radars are complementary. Thus here is a good opportunity to utilize wind data from the two different radars through a year, which then enables us to cover a wide height range (approximately 70-120 km). This wide range gives us good opportunities to examine the mesosphere/lower thermosphere coupling process based on observational data.

1 Introduction

The atmospheric dynamics in the lower thermosphere and upper mesosphere is strongly controlled by atmospheric waves such as gravity, tidal, and planetary waves which mainly propagate upward from below. In these regions, substantial amount of gravity wave energy must reach the lower thermosphere from below [see e.g., *Hocking*, 1996]. *Hocking* [1996] pointed out the importance of comparative studies of winds at different heights in these regions. *Nozawa and Brekke* [1999a,b] studied seasonal and solar cycle variations of the mean and tidal winds in the lower thermosphere between 95 and 119 km. They showed, for example, a strong eastward mean wind ($\sim$40 m s^{-1}) in summer maximizing at 101 km, and suggested that this feature in the altitude wind profile is produced by gravity wave drag. It is, however, impossible to prove this idea quantitatively by using observational data in the lower thermosphere only. To understand the lower thermospheric dynamics more completely, therefore, it is necessary to make simultaneous observations of the wind both in the lower thermosphere and the upper mesosphere. There are various techniques to observe (or derive) wind velocity in the lower thermosphere as well as the mesosphere. Regarding the ground-based observations, radar techniques are generally superior to optical techniques in terms of tidal studies because radars can make measurement continuously for 24 hours a day with relatively good height resolution ($\sim$3 km or so). However, no radar techniques can be applied to derive wind measurements simultaneously both in the upper mesosphere and the lower thermosphere. For example, a medium frequency (MF) radar can observe winds regularly from $\sim$60 km to $\sim$100 km [e.g., *Reid*, 1996], while an incoherent scatter (IS) radar can derive wind velocity from $\sim$90 km to $\sim$120 km [e.g., *Nozawa and Brekke*, 1999a,b]. Therefore, we need to utilize at least two radars to cover the whole region, one of the most unique regions in the earth's atmosphere.

In recent years several papers [e.g., *Hocking*, 1997; *Reid*, 1996] have questioned the validity of MF radar observations above 90 km. Problems, for example, due to receiver saturation, strong reflection from E-layer and group retardation are pointed out. *Namboothiri et al.* [1993] investigated the difference between real and virtual heights for 2.2 MHz radio wave using standard electron density profiles from rocket observations as well as the International Reference Ionosphere (IRI) model [*Rawer et al.*, 1978] for application to the Saskatoon (52°N, 107°W) MF radar. They concluded that in winter no corrections for group retardation are necessary up to 100 km height, while during summer such corrections are necessary above 95-97 km. *Hall* [1998] estimated group retardation for 2.8 MHz radio wave by using the empirical electron density model [*Friedrich and Torkar*, 1981] which is based on a substantial number of rocket soundings from the Andøya Rocket Range (69°N, 16°E) situated only 120 km to the west of Tromsø. *Hall* [1998] discussed virtual to real height correction for the Tromsø (69.58°N, 19.23°E) MF radar (ν=2.8 MHz) [e.g., *Hall*, 2001]. *Hall* [1998] supported the conclusions by *Namboothiri et al.* [1993] and noted that the upper limit was 94 km for reliable observations for the Tromsø MF radar. These two studies, however, were based on models for electron density profiles. At high latitudes during geomagnetically disturbed periods particles originating in the magnetosphere penetrate into the lower thermosphere and even into the upper mesosphere, lowering the total reflection height of 2-3 MHz radio wave down to 100 km on some occasions and thereby causing severe group retardation effects below 100 km.

Hines et al. [1993] made a comprehensive comparison of winds in the mesosphere and lower thermosphere obtained during the Arecibo Initiative in Dynamics of the Atmosphere (AIDA) '89 campaign. They compared 14 profiles of wind data derived by the MF/HF partial-reflection technique with those from the Arecibo IS radar at heights of 65-95 km and meteor radar at heights of 80-100 km colocated in Arecibo (18°N, 67°W). They concluded that wind velocities derived by the MF/HF radar are good representations of ambient winds only up to about 80 km. Contrary to the conclusion by *Hines et al.* [1993], *Turek et al.* [1995] examined 208 profiles of the AIDA campaign wind data and concluded that both the imaging Doppler interferometer (IDI) and the Arecibo IS radar winds are normally distributed over the height range from 70 to 97 km and the IDI wind compares well with winds derived from other MF/HF radar wind measuring techniques such as full correlation analysis (FCA) and three dimensional interferometry analysis (INT). Furthermore *Turek et al.* [1998] again compared the Arecibo IS radar wind data with MF/HF wind data derived from several techniques (IDI, FCA and INT). They found in general very good agreement (within 15 m s^{-1}) up to 90 km between the zonal winds determined from all the techniques. Furthermore, in the last decade comparisons of winds from MF radar and ground-based optical instruments has been widely carried out [e.g., *Phillips et al.*, 1994; *Manson et al.*, 1996; *Meek et al.*, 1997], and generally good agreement was illustrated at a height of $\sim$97 km. For example, *Manson et al.* [1996] made comparison between instantaneous measurements made at Saskatoon (52°N, 107°W) using the colocated MF radars and Fabry-Perot interferometer (FPI) instruments over 1988-1992. They found good agreement (the speed ratios of winds from two instruments are close to unity; the vector angle differences are centered about 0°, with medians less than +5°) on the climatological comparison. *Meek et al.* [1997] made comparisons of winds derived from several instruments such as Canadian prairie MF radars, FPI, and the Upper Atmosphere Research Satellite (UARS) high resolution Doppler imager (HRDI) systems. On the statistical comparisons of hourly mean winds obtained over 1988-1992 for the Saskatoon and Calgary MF radars and FPIs (OH and 557.7 nm green line), they found excellent agreement (e.g., speed ratio as well as phase difference are close to 1.0 and 0 degree, respectively) for Saskatoon at two heights of $\sim$88 km and $\sim$98 km. However, *Meek et al.* [1997] also showed some discrepancy that the Sylvan Lake MF radar found values larger than the FPI (88 km layer) by a factor of 1.2 and Saskatoon MF radar speeds were less than HRDI values by factors 0.7-0.85. *Burrage et al.* [1996] made comparisons of winds derived from the HRDI on UARS and MF radars, finding reasonable agreement (when considering the very different sampling conditions of a remote sensing technique and a localized system) between 60 and 95 km, but discrepancy in the relative magnitudes above 95 km. Therefore, the validity of MF radar wind measurements above $\sim$90 km is still an open question.

At high latitude, there are only few studies made to compare winds derived by MF radar and other instruments. *Manson et al.* [1992] made the first comparison of winds at high latitudes based on the Tromsø MF radar, VHF radar at Andenes, rockets launched from Andenes, and the European Incoherent Scatter (EISCAT) [*Folkestad et al.*, 1983; *Rishbeth and Williams*, 1985] Common Program One (CP-1) data for six short IS campaigns during 1987-1989. MF radar winds were shown to be ~0.65 of those from the Andness data (rocket and VHF radar). However, until 1988 neutral wind velocities could not be derived below 100 km using the CP-1 mode with the tristatic method [e.g., *Williams et al.*, 1994]. Thus, during these campaigns too few events occurred for comparing winds derived from simultaneous EISCAT and MF radar data. They concluded, however, that once modified in amplitudes by the 1.5 factor there was, in general, very favorable agreement in phase and amplitude between the EISCAT and MF radars (e.g., in case of June 1987, the semidiurnal amplitude and phase near 100 km derived from two radars are very similar: the differences of amplitude and phase are a few m s^{-1} and ~1 hour, respectively). Differences were attributed to low signal to noise (S/N) conditions for either radar, lack of complete data overlap, and strong wind gradients in the lower thermosphere.

In this paper, we have made comparisons of winds derived by the these two radars which are colocated in Ramfjordmoen, nearby Tromsø in Norway. Section 2 is devoted to descriptions of these radars and their methods of observations, and comparisons of winds obtained simultaneously by the both radars are presented in Section 3. In Section 4, mean and tidal winds derived from 1-month averaged wind data obtained by the Tromsø MF radar for 1999 are compared with those of the statistical study by *Nozawa and Brekke* [1999a]. Discussions are given in Section 5, and summary and conclusions will be given in Section 6.

2 Observational methods

2.1 The Tromsø MF radar

The Tromsø MF radar has been in operation for more than 10 years in a spaced antenna (wind measuring) mode. A recent specification of this radar can be found in *Hall* [2001]. In summer 1998 its transmitter antenna system was refurbished, and in early November in 1998 a new solid state transmitter (50 kW) was installed. After these improvements, the Tromsø MF radar has furnished wind data more routinely in the height region from ~70 to ~100 km (in virtual height) on a diurnal basis. Since then, this radar system has been operated under collaboration of the Universities of Saskatchewan and Tromsø and Nagoya University. We have assembled all data obtained by the MF radar in 1999. The time resolution was 5 min until February 15, 1999 and since then it has been 2 min. The Tromsø MF radar receives signals (partially) reflected in the mesosphere and lower thermosphere with thirty-two 3 km wide gates starting at 40 km, and its analysis is based on the Saskatoon Full Correlation Analysis (FCA) technique of Spaced Antenna (SA) data [*Briggs*, 1984; *Meek*, 1980]. The SA technique is most commonly applied to MF radars [see e.g., *Reid*, 1996]. The Saskatoon FCA technique involves sampling coherent fading signals simultaneously at three or more, non-colinear, spaced antennas at a set of fixed height gates (40-133 km with a step of 3 km). The resulting complex time sequences at each height are correlated (cross and auto) and fitted with a Gaussian correlation function [*Meek*, 1980] which includes ground-pattern spatial anisotropy and decay terms, as well as the velocity vector. The ground pattern is assumed to be moving twice the speed of the scattering irregularities. Because the system is coherent, doppler velocity and angle of arriving phase fronts are also calculated. Generally, errors of wind velocities from the MF radar based on the FCA method is estimated to be less than $\sim$10 m s^{-1} [see *Meek and Manson*, 2001].

The MF radar operates continually. For the comparative analysis of simultaneously obtained MF and EISCAT radar data, we use the MF wind data with 2 min or 5 min resolution, while for statistical comparisons, we calculated monthly averages. We only use the MF wind data obtained below the height where the receiving power maximizes [see e.g., *Namboothiri et al.*, 1994]. Owing to this criterion for acceptance of wind data, the number of data is significantly reduced above 90 km and no wind data remain at and above 106 km over the year 1999. In order to secure data reliability, this criterion is necessary and commonly applied to the Saskatoon MF radar data in these days. Hereafter we refer to this as "MF criterion". Based on the 1-month averaged data obtained by the Tromsø MF radar, we have derived mean wind and diurnal and semidiurnal amplitudes and phases for each month in 1999. The tidal components are derived by using the Lomb-Scargle periodogram method which is based on least-squares frequency analysis of unequally spaced data [cf. *Hocke*, 1998].

2.2 The EISCAT UHF radar

Among the EISCAT CP modes [e.g., *Collis*, 1995], CP-1 and CP-2 modes have been used for deriving neutral wind velocity vectors in the E region. In the CP-1 mode the line of sight of the combined transmitter and receiver antenna is fixed along the magnetic field line at Tromsø at an elevation angle of 77.5°, while in the CP-2 mode the line of sight of the antenna is pointed into four consecutive positions with a dwell time of $\sim$1 min in each position, resulting in a full cycle time of the antenna of 6 min. There exist three observational methods to derive neutral wind velocities such as the tristatic method (applied to earlier CP-1 mode), the monostatic method (CP-2 mode), and the field-aligned method (mainly CP-1 mode) [see *Williams et al.*, 1994; *Nozawa and Brekke*, 2000]. In this study, we have derived meridional wind velocities from data obtained by the CP-1 mode based on the field-aligned method. On the other hand, from data obtained by the CP-2 mode (monostatic method) we have derived both meridional and zonal wind velocities. Here we briefly describe the ways to derive neutral wind velocities from ion velocity measurements [see details in *Williams et al.*, 1994].

Like other IS radars, the EISCAT radar measures ion velocities, not the wind velocity directly, in the ionosphere. In the lower thermosphere due to the strong coupling between ions and neutrals one can derive the neutral wind velocity vector ($\mathbf{u}$) using both measured ion velocity ($\mathbf{v}$) in the E region and derived E-field ($\mathbf{E}$) from F region ion velocity measurement by using the following equation [*Rino et al.*, 1977]:

$$\mathbf{u} = \mathbf{v} - \frac{\Omega_i}{|\mathbf{B}|\nu_{in}}(\mathbf{E} + \mathbf{v} \times \mathbf{B}) \tag{1}$$

where $\mathbf{B}$ is the earth's magnetic field, ν_{in} is ion-neutral collision frequency, and Ω_i is ion gyrofrequency. We base our analysis on an International Geomagnetic Reference Field (IGRF) model (IAGA Division I working Group 1, 1987) and a formula for the ion-neutral collision frequency introduced by *Schunk and Walker* [1973]. The model neutral atmosphere is given by MSIS86 model [*Hedin*, 1987]. This method (equation 1) was firstly introduced by *Brekke et al.* [1973], evaluated by *Comfort et al.* [1976], improved by *Rino et al.* [1977]. From this technique, which can be applied to data obtained by both the tristatic and monostatic methods, one can derive tri-dimensional neutral wind velocity vectors in the lower thermosphere [*Nozawa and Brekke*, 1999a].

Another way is the field-aligned method. This method takes the component of the ion velocity vector ($v_{//}$) measured along the local magnetic field line, which is at an inclination of I = 77.5° and azimuth of 182.6° (counting from north to east) at Tromsø. In the lower thermosphere, where ion diffusion velocity along the magnetic field line is significantly small (less than 1 m s^{-1}) [e.g., *Kofman et al.*, 1996], the ion motion along the magnetic field is independent of electric field and dominated by neutral wind drag. Thus the field-aligned ion velocity ($v_{//}$) is expressed as follows:

$$v_{//} = -u_h \cos I + u_v \sin I \tag{2}$$

where u_h and u_v are the horizontal and vertical neutral wind velocity, respectively. It follows that if the vertical velocity of the neutral wind (u_v) is considerably smaller than u_h cot$I$, then the horizontal neutral wind velocity is given by [*Williams et al.*, 1994]:

$$u_h = -v_{//}/\cos I \tag{3}$$

At the EISCAT Tromsø site, where the local magnetic field is directed almost south, the relative contributing factor of zonal component to $\mathbf{u}_h$, compared to that of the meridional component, is 0.045 (= tan (182.6°)). Thus, the meridional component of neutral wind can be determined in this way from eq. (3). Although the field-aligned method can furnish only meridional wind velocity, it has proved very effective in identifying tidal modes over the height range 100-160 km [*Virdi et al.*, 1986; *Huuskonen et al.*, 1991] and the best one among the methods for the meridional component [*Nozawa and Brekke*, 2000]. Comparisons of derived winds from different modes and/or methods were conducted on a statistical basis [*Nozawa and Brekke*, 1999a] and on a case study basis [*Nozawa and Brekke*, 2000], and showed that the derived winds agreed reasonably well in terms of longer temporal variations (>∼3 hours). *Nozawa and Brekke* [2000], however, showed that derived wind data by the monostatic method scatters with time compared to those derived from the tristatic method and the field-aligned method.

Table 1 summarizes data obtained by the CP-1 and CP-2 modes in 1999, which are used in this study. In addition, averaged system noise temperature as well as averaged transmitter power for each experiment are listed in Table 1 to give a general idea of conditions of the EISCAT UHF radar. As Table 1 shows, there are more than 20 days when temporal wind variations can be directly compared with the MF radar winds. On the other hand, *Nozawa and Brekke* [1999a,b] studied seasonal and solar cycle variations based on 56 days of EISCAT CP-1 and CP-2 data obtained from 1987 to 1996. We are utilizing the same data base for computing the mean wind and diurnal and semidiurnal components in this work.

It should be pointed out that for IS radars (based on the normal IS theory [e.g., *Schlegel*, 1995]) measurement is less reliable below 100 km than above. This is because the electron density decreases significantly below 100 km. In particular, wind velocities derived by the monostatic method unfortunately involves rather large uncertainty. In contrast, the field-aligned method is more reliable than the monostatic method, since longer integration time for each IS spectrum can be applied [see *Nozawa and Brekke*, 2000].

3 Comparisons of wind velocities based on simultaneous data

Here we present comparisons of temporal wind variations based on simultaneous observations by the EISCAT and MF radars co-located in Tromsø. Table 2 summarizes geomagnetic activity (3-h Kp and daily Ap indices) and the solar 10.7 cm flux ($F10.7$ index) for these events. The dates for comparison are distributed from quiet to moderate geomagnetic activities under high solar activity conditions.

We mention at first horizontal resolutions of EISCAT and MF radars. The EISCAT UHF radar observes a column volume (diameter = ∼1 km and height = ∼3 km) in the E region. In the CP-2 mode, to derive tri-dimensional ion velocity vectors, we need to combine three line of sight measurements made at three column volumes which are 30∼50 km apart from each other in the E region. On the other hand, for the MF radar, it is a bit difficult to determine the volume like the EISCAT radar. If angular spread is estimated from FCA ground pattern scales, values for scattering volume extent of the order of 10 km are found for the Tromsø MF radar. We think these differences of the resolutions introduce disagreement of wind velocities for short-time periods into the wind comparisons, but the variations of winds should be similar for a few hours or longer period (especially for mean and tidal winds).

On the basis of the propagation theory of error, each error of the neutral wind velocity derived by the EISCAT radar is calculated from errors determined originally by the IS spectrum analysis. Concerning error bars of tidal components, we firstly calculate a standard deviation for each point of time when wind data are averaged. Then we calculate an averaged root-mean-square of the standard deviations for each averaged data set such as seasonal averaged data for EISCAT radar and monthly averaged data for MF radar. Using the averaged root-mean-square of standard deviations (we refer to this as averaged standard deviation, hereafter) we finally estimate standard deviations for all mean and tidal components from EISCAT and MF radars and shown them as error bars (1σ) in corresponding figures. The way of calculation is as follows:

$$\sigma_{A_{mean}} = \sqrt{\frac{\sigma^2}{n}} \tag{4}$$

$$\sigma_{A_i} = \sqrt{2\frac{\sigma^2}{n}} \tag{5}$$

$$\sigma_{\phi_i} = \sqrt{2\frac{\sigma^2}{{A_i}^2 n}} \tag{6}$$

where $\sigma_{A_{mean}}$, σ_{A_i}, $\sigma_{\phi i}$ are standard deviations for mean wind, tidal amplitude and phase (diurnal and semidiurnal components), respectively, and n is the number of data points with time, σ is the averaged standard deviation derived from $\sigma = \sqrt{\Sigma \frac{\sigma_k^2}{n}}$, and σ_k is the standard deviation at point k (of time). A detailed description of this calculation is presented in *Nozawa and Brekke* [1995].

It is important for radars using medium frequency to identify where transmitted radio wave is (partially) reflected. Generally, electron density increases with increasing height in the mesosphere and lower thermosphere, group delay of the radio wave becomes more significant with increasing height. To clarify the observed height for MF radar wind data, we present differences between true and virtual heights calculated using electron density profiles obtained simultaneously by the EISCAT radar. The EISCAT electron density used in this study is calibrated by using dynasonde "fof2" data. The difference dh in km as a function of height is give by:

$$dh(z) = z - \int_0^z \sqrt{1 - \frac{80.5 N_e(z)}{{f_r}^2}} dz \tag{7}$$

where z is the true height in km, N_e is the electron density in m^{-3} as a function of (true) height given by EISCAT observations, and f_r (= 2.8 MHz in this case) is the radar frequency in Hz [*Hall*, 1998; *Namboothiri et al.*, 1993]. EISCAT CP-1 and CP-2 modes make measurements of electron density from 62.3 km to ~582 km. Thus, in practice, the difference is calculated by:

$$\Delta h(z) = z - (62.3 + \sum_{62.3}^{z} \sqrt{1 - \frac{80.5 N_e(z)}{{f_r}^2}} \Delta z) \tag{8}$$

where Δz is a gate interval which is 3.15 km from 62 to ~269 km for both short pulse (power profile) and alternating code (ACF) and 22.5 km from ~142 km to ~582 km for long pulse [*Collis*, 1995]. Ambiguity of EISCAT electron density measurement is expected to be about 10% or so. When electron density is higher ($> 3\times10^{10}$ m^{-3}), its ambiguity of measurement becomes lower. Therefore we estimate an ambiguity in Δh due to 10 % ambiguity in electron density. Although we can derive the difference between true and virtual heights by using Eq. (8), we present MF radar wind data as a function of virtual height in figures to avoid possible confusions. Readers should keep in mind that there would be difference between true and virtual heights (especially above 90 km) when comparing the wind data. Also, it should be noted that there is a height difference up to 1 km between the centers of each gate of both radars. Since their height resolutions are about 3 km, this mismatch of the center gate probably does not cause any serious problems with the comparison. On instantaneous wind comparison, EISCAT wind data with errors greater than 50 m s^{-1} are always rejected.

As shown in Table 1, on case study basis we made comparison of wind data derived with the EISCAT and MF radars for five cases. Here we present three cases from each season (winter, summer and equinox) in detail, and for the other two cases brief summaries are given.

3.1 Case 1: February 9-12, 1999

Figure 1a shows meridional wind variations as a function of universal time obtained by the EISCAT CP-1 mode (plus) from 10 UT on February 9 to 16 UT on February 12, 1999 for four heights from 90 km to 99 km. The wind data was derived by the field-aligned method with post-integration time of 5 min. Meridional and zonal components of the electric field as well as 3-h Kp index are also shown in the upper two panels and in the bottom panel, respectively, for examining possible relationship between winds below 100 km and geomagnetic activity. Corresponding wind data (open circle) obtained by the Tromsø MF radar are shown for comparison.

Geomagnetic activity was generally quiet ($Kp < 3$) on February 9 and 10, and was moderate ($Kp =\sim 3-4$) on February 11 and 12. The amplitude of zonal component of the electric field was small (less than ~10 mV m^{-1}) almost over the period except between approximately 0100 and 0300 UT on February 12 when it was enhanced up to ~26 mV m^{-1} eastward. Meridional component of the electric field showed enhancements (up to ~60 mV m^{-1} northward) between approximately 1700 and 2300 UT on February 9 and 10. On February 11 it was again enhanced to be ~70 mV m^{-1} northward at around 1430 UT, and turned from northward to southward at around 2200 UT. The electric field amplitude reached ~80 mV m^{-1} southward at around 0200 UT, and the electric field again turned northward and was enhanced between 1100 and 1500 UT on February 12. When comparing MF radar wind data with the strength of the electric field during the periods, no clear relationship can be found between them. For example, between 0100 and 0300 UT on February 12, when E-field is strongly enhanced, the variation of MF radar wind at 90 km did not show large departure from that for other periods, and the MF radar wind velocity (except for one) ranged from -50 to 50 m s^{-1} over the height region.

Although data gaps often occurred, it can be identified that the EISCAT wind data ranged from approximately -150 to 150 m s^{-1} and showed semidiurnal variation over the height region. At 90 km, variation of the MF radar wind data showed small fluctuation, but it tended to fluctuate more with height increasing. In addition, the number of MF radar wind data tended to decrease with height increasing. When comparing variations of wind velocity derived by EISCAT and MF radars over the height region from 90 to 99 km, in general smoother variation in time is seen for the MF radar wind data than for the EISCAT radar data. The amplitude of MF radar wind data varied in smaller range than that of the EISCAT wind data. EISCAT wind derivations succeed during sunlit-time as well as for particle precipitation periods, while MF radar wind data obtained for particle precipitation periods are often discarded by the MF criterion (see section 2.1). Between 90 and 96 km, wind velocities from two radars appear to correspond well, and at 99 km no strong departures of the wind velocities between EISCAT and MF radar data can be found in this case.

Figure 1b illustrates differences between real and virtual heights for 2.8 MHz radio wave for the experiment period between 84 and 99 km. When total reflection occurs at and below the height, a value of -1 is shown. Furthermore, at the bottom of Figure 1b, temporal variation of height integrated Hall conductivity (conductance) calculated according to the work by *Brekke and Hall* [1988] is shown, since the conductance could be a good indicator of particle precipitation. Generally, the differences are small up to 90 km. When no particle precipitation occurred, the differences were less than 3 km (= 1 gate interval) up to 105 km over the experiment period.

However, MF wind data was affected by group retardation or total reflection due to particle precipitation for almost half of time at and above 99 km, and below 99 km such influence can be often seen. Even at 90 km, the influence is found for time periods between 0030 UT and 0200 UT as well as 0600 UT and 1000 UT on February 11 and between 0600 UT and 1300 UT on February 12, indicating care is necessary for MF radar wind data above 90 km at high latitude station.

3.2 Case 2: July 1-9, 1999

This EISCAT experiment was conducted partly as a CP and partly as a special program [*Collis*, 1995] during this period. Its observational mode was identical to the CP-2 mode in which tri-dimensional wind data are furnished at every 6 min. The EISCAT data, unfortunately, were affected badly by the high system noise temperature (~156 K), approximately twice as high as normal. This makes the comparison more difficult. Figures 2a and 2b show comparisons of the winds for the meridional and zonal components, respectively. Same to Figure 1a, variations of meridional and zonal components of the electric field and 3-h Kp index are illustrated at the top 2 panels and the bottom, respectively. It was geomagnetically quiet (Kp <3) almost whole the experiment period except for a day of July 2. The amplitude of the meridional component of the electric field was enhanced northward (up to ~40 mV m^{-1}) between approximately 1600 and 2100 UT on July 2 and between approximately 1300 and 2000 UT on July 8. The electric field is also enhanced southward (up to ~30 mV m^{-1}) between approximately 2200 and 0300 UT on July 5-6 and between approximately 2100 and 0200 UT on July 8-9. For other period the amplitude of electric field was generally small (less than 10 mV m^{-1}). During the period when the electric field was enhanced, MF radar wind velocity of the horizontal components over the height region ranged approximately between -50 and 50 m s^{-1} except period between 1600 and 2100 UT on July 2 when the meridional wind data at 90 km height ranges approximately 20 m s^{-1} to -100 m s^{-1}. Therefore, we can conclude from this event that no clear relationship can be found between strength of electric field and MF radar wind velocity.

Very low number of wind data is obtained by EISCAT at and below 92 km in this experiment. Although the wind data with their error values of more than 50 m s^{-1} were eliminated, the EISCAT wind data exhibit much scatter over the height region. This is probably due to several causes such as low electron density below 100 km, high system noise temperature, and shortcoming of the monostatic method [*Nozawa and Brekke, 2000*]. For the MF radar wind data the number of values is high compared with that in other seasons and they show smoother variations with time than those of EISCAT wind data. At 89 km MF radar wind data have almost no data gaps, while wind derivation by EISCAT failed for almost all the period. The number of the MF radar wind data obtained decreases with height increasing, and then at 98 km the MF radar wind data were obtained for one third of the experiment period, but still they showed smaller scatter than the EISCAT wind data. These fact indicate that poor agreement for this case between 89 and 98 km for the horizontal component is primarily attributed to the higher ambiguity of the EISCAT wind measurements due to the high system noise temperature.

Figure 2c shows the difference between true and virtual heights as well as the Hall conductance. From the variation of the Hall conductance, it can be identified during night time particle precipitation often occurred, resulting the 2.8 MHz radio wave was often totally reflected at and below 93 km. When no particle precipitation occurred, the difference is about half a gate interval ($\sim$1.5 km) below 90 km almost all the time. However, the difference becomes larger with increasing height. At and above 96 km, the difference shows a clear diurnal variation likely due to solar insolation. Sometimes total reflection occurred at 99 km regardless of particle precipitation. These facts imply that although the number of values is high, special care is needed for the MF radar wind data above 90 km in summer under high solar activity conditions at high latitudes.

3.3 Case 3: September 15-17, 1999

Figure 3a compares meridional wind data obtained by the EISCAT and MF radars between 90 and 99 km for equinox season. Again variations of the meridional and zonal components of the electric field and 3-h Kp index are presented altogether with the wind data. The meridional component of the electric field varied in time largely from about 0000 UT on September 16 and to 0300 UT on September 17, and it ranged from approximately -80 mV m^{-1} and 70 mV m^{-1}. It was small between 0300 and 1200 UT on September 17 and again increased up to about 50 mV m^{-1} northward at the end of the experiment. The zonal component of the electric field also showed enhancements (up to 30 mV m^{-1} eastward) between about 0500 UT and 0600 UT on September 16 and between about 0200 UT and 0300 UT on September 17. The 3-h Kp index was high (up to 6) on September 16. Again, no clear relationship between MF radar wind velocity and strength of the electric field can be found. In this case, the number of MF wind data is relatively low over the height region, and until 07:40 UT on September 16, no data were obtained by the MF radar due to a system failure. The EISCAT wind data show a clear semidiurnal variation ranging from approximately -130 to 130 m s^{-1} over the height region with a good data coverage. MF radar wind data obtained for particle precipitation periods appear to be discarded due to the MF criterion. Although the number of the MF radar wind data is low, it can be identified that both wind data correspond relatively well between 90 and 96 km. At some times MF radar wind velocity significantly departed from that of the EISCAT wind velocity at 99 km between 0400 and 0500 UT on September 17, but these are likely due to effects of group retardation or total reflection as shown in Figure 3b (i.e., true height of those MF wind data should be lower).

Figure 3b shows the difference between true and virtual height for the 2.8 MHz radio wave between 84 and 99 km as well as the Hall conductance during the experiment period. From a plot of the Hall conductance, it can be identified particle precipitation almost always occurred on September 16 and 17, resulting group delay as well as total reflection occurred for almost all the time period. On September 15, during no particle precipitation or soft particle precipitation (no particle reached to the E region) period, the difference is less than 3 km up to 105 km. However, on September 16 and 17, during the particle precipitation period, the total reflection height went down below 90 km (down to 84 km or even below) on some occasions. During the period between 0300 and 0900 UT on September 16, total reflection occurred below 90 km almost all the time. On September 17 at 96 km and 99 km, the difference ranged from 1.5 km to 9 km (or more), and it fluctuated with time. These facts again imply that special care is necessary with the MF radar wind data obtained at high latitude station.

3.4 Other two cases

We have also made comparison for other periods such as March 8-12 and October 12-15, 1999. In the case of March 8-12, both the EISCAT and MF radar wind data have long data gaps. Owing to system failures, low numbers of the wind data were assembled for March 1999. Although the occasions for comparison are few, at least no significant departure of wind velocity between two radars are identified. Geomagnetic activity was high on March 9 and 10, when the radio wave was affected by group retardation or total reflection for more than one third of the observing period at 90 km. In the case of October 12-15, the number of MF radar wind derivation is very low at and above 90 km. Effects due to likely particle precipitation can be seen at lower height of 84 km. At and above 90 km the MF radar suffers from group retardation or total reflection for more than 40 % of the period.

Finally, it should be pointed out that over all the instantaneous wind comparisons presented in this section, it is likely that the MF criterion lead to data rejection for periods of strong group retardation or total reflection occurring due to particle precipitation.

4 Comparison of mean and tidal winds

Based on 1-month-averaged wind data obtained by the Tromsø MF radar, we have derived mean winds as well as amplitudes and phases of diurnal and semidiurnal components for each month for 11 heights from 70 to 100 km (in virtual height). Owing to low numbers of wind data, these are derived from 70 to 91 km for March and from 70 to 94 km for October. In 1999 the solar activity was high (see Table 2), and the averaged value of $F_{10.7}$ index for 1999 is $\sim 160 \times 10^{-22}$ W m^{-2} Hz^{-1}. *Nozawa and Brekke* [1999a] collected 56 days of EISCAT CP-1 and CP-2 wind data between 95 and 119 km obtained under geomagnetically quiet conditions ($Ap < 16$) over a solar cycle from 1987 to 1996. The averaged $F_{10.7}$ index over those 56 days was $\sim 142 \times 10^{-22}$ W m^{-2} Hz^{-1}. In order to clarify EISCAT and MF winds for climatological study, here we compare these statistical EISCAT wind data with the MF radar wind data in terms of mean wind and tidal (diurnal and semidiurnal) amplitudes and phases, and then we examine the upper limit for reliable wind observations of the MF radar from a statistical viewpoint. In the study of *Nozawa and Brekke* [1999a], the wind data were sorted into three seasons such as summer (May 3 - August 18), equinox (February 24 - April 14, and August 30 - October 20) and winter (October 21 - February 20).

Figure 4a compares meridional and zonal mean winds derived by the EISCAT and MF radars for each of these seasons. For the height of the MF radar data no correction to the group retardation is made, meaning the wind values derived by the MF radar are shown as a function of virtual height. Readers should keep in mind that differences between true and virtual height is, generally, 3-6 km above 90 km for the summer data. Concerning the altitude profile obtained by the MF radar, month-to-month variation in the same season appears less prominent except for the zonal component in the equinox season when considerable changes occur in the tidal forcing as well as back ground wind and temperature [e.g., *Manson et al.*, 1999]. Relatively good agreement (differences are less than $\sim$10 m s^{-1}) is found in the meridional component for equinox. The altitude profile of the zonal mean wind in summer, which is the most structured in altitude, obtained by the MF radar is fairly constant with height above 91 km. In summer the electron density in the ionosphere is high, and this assures more reliable IS radar measurements than other seasons. In contrast, high electron density causes group retardation and total reflection below 100 km as shown in Figure 2c. Therefore, it appears that reasonable variations as a function of height are seen up to 91 km for the MF radar wind data. In winter, the tendency that the eastward wind amplitude reduces gradually with increasing height appears to be reasonable, but at 95 km the difference of more than 20 m s^{-1} is found between the winds obtained by the different radars. It is worth noting that winds modeled by Canadian Middle Atmosphere Model (CMAM) [e.g. *McLandress*, 1998] and provided from meteor wind radar at similar latitudes (Mitchell, private communication) do show the reversal height near 100 km. The EISCAT winds below 100 km thus might be questionable, but we cannot neglect the fact that the MF radar wind data above 90 km appear more noisy than those below 90 km.

Figure 4b illustrates altitude profiles of the diurnal amplitudes and phases derived by EISCAT and MF radars. Fairly good agreement (the difference being less than $\sim$10 m s^{-1}) is seen for meridional and zonal amplitudes except for winter where large differences ($\sim$30 m s^{-1}) exist at 95 and 98 km for the zonal component. The EISCAT values for winter below 100 km are much larger than those at 100 km, suggesting a problematic result. These cannot be separated from the large mean winds in Figure 4a. Concerning phases, good agreement ($<\sim$3 hours) is found for meridional components for summer and equinox. The meridional phase values obtained in January and December exhibit large (up to $\sim$11 hours) differences between them. The difference of the zonal phase component between EISCAT and MF radar data is large for summer ($<\sim$10 hours) (again perhaps related to the mean wind values in Figure 4a), while relatively good agreement ($<\sim$6 hours) is found for equinox and winter (except for January data) considering their large change with month. The winter values are again noisy in appearance.

Figure 4c compares semidiurnal amplitudes and phases. Significant differences for summer and winter are identified in the amplitudes. Especially in summer, are there large ($>$20 m s^{-1}) disagreements at and above 95 km. Concerning phases, they appear to be in fairly good agreement ($<\sim$3 hours) for almost all the seasons. However, the phase values of the meridional component in summer months tend to be constant with height at and above 91 km in June and between 88 and 94 km in July, implying the occurrence of group retardation.

Finally, if we combine altitude profiles of winds derived from the MF radar between 70 and 91 km and the EISCAT radar between 95 and 120 km, it seems like that they can smoothly connect to each other (except for zonal mean wind and diurnal components both in winer).

This feature is very encouraging for future studies.

5 Discussion

In this paper, we have shown comparisons of winds between 90 and 100 km derived from two radars: the EISCAT and MF radars. The most important fact is to examine how well we can utilize the wind data obtained from the different radars in order to cover a wide height range, which enable a detailed research on the mesosphere/lower-thermosphere coupling process on an observational basis. Generally, IS radar measurement under geomagnetically quiet conditions is reliable between ∼100 and 120 km regardless of season, while MF radar measurements work well between approximately 70 and 90 km. As shown before, however, the wind measurements (or derivations) from the two radars are relatively difficult between 90 and 100 km especially on an instantaneous basis. In summer, the high ionization assures the reliable IS radar measurement even below 100 km, while it causes group retardation or total reflection of the MF radio wave. Thus, the IS radar measurement is likely more reliable than the MF radar measurement between ∼90 and 100 km in summer. On the other hand, in winter the IS radar measurement is difficult below 100 km due to low electron density, especially for night-time. In contrast, low electron density allow the MF radio wave to penetrate to the E region heights.

In Figures 1a, 2a, 2b, and 3a, in addition to temporal wind variations, we also present the 3-h Kp index to examine the effects of geomagnetic disturbance on the the MF radar wind measurement as well as the relationship with the occurrence of the group retardation and total reflection. From Table 2, the daily Ap index varied between 3 and 46 for the events studied. As we have shown, under these conditions, an effect of auroral particle precipitation sometimes penetrated into the D region. In the case of February 1999, the particle precipitation occurred when Kp was less than 2. Since particle precipitation and Kp index show no direct relationship, it is difficult to evaluate the validity of wind velocity from the MF radar by using only Kp index. However, since we use the wind data obtained below the height where receiving power maximizes (MF criterion), the wind data which would suffer from possible contamination due to particle precipitation are likely rejected.

It is clearly shown in Figure 2c that in summer when no particle precipitation occurred, the group retardation became significant at and above 93 km. At 99 km, the difference ranged from 1 to 6 km. From these facts, we can conclude that in summer under high solar activity conditions the upper limit for reliable MF radar wind observations is 91 km at this latitude. This conclusion agrees well with that of *Hall* [1998], who calculated the group retardation based on an the empirical electron density model. Furthermore, our conclusion is more critical to the virtual height correction than that of *Namboothiri et al.* [1993]. For equinox season, when no particle precipitation occurred, it appears that the MF radar can furnish reliable wind data up to ∼96 km under high solar activity conditions. However, when the particle precipitation occurs, the wind measurement at and above 90 km is no longer reliable. In winter, during periods at no particle precipitation, the effect of group retardation is small (less than 1 gate interval) up to 105 km. Fairly good agreement is found between winds by the two methods up to ∼96 km when no particle precipitation occurred as shown Figure 1a. Therefore, it is concluded that concerning instantaneous wind measurement during no particle precipitation period, the MF radar can furnish reliable data up to ~96 km except for summer. During periods of particle precipitation, it is unlikely the MF wind measurements are reliable above ~91 km.

In the statistical study, as shown in Figures 4a, 4b and 4c some discrepancies between the EISCAT and MF radar winds are found at and above 95 km. By comparing altitude profiles of the zonal mean wind as well as the semidiurnal amplitude, it appears that MF radar wind measurements are questionable above ~91 km in summer, although the semidiurnal phase values from two radars are in good agreement (difference being $<\sim$3 hours) at and above 95 km. This good agreement of the phase should be a coincidence. These amplitude disagreements could be caused by the effect of group retardation and/or an effect such as strong reflection from the *E* layer as pointed by *Hocking* [1997]. For equinox, considering its variability from month-to-month as well as the limited number of EISCAT wind data (averaged spring and fall data together), the agreement of mean and semidiurnal components can be concluded to be relatively good. The EISCAT tidal wind parameters are often quite variable below 100 km.

On the other hand, the disagreement of altitude profiles of the winter mean wind can suggest an important fact. Eastward mean wind amplitude derived by the MF radar at 70 km is about 20 m s^{-1} and it decreases with increasing height and close to zero above 90 km. In contrast, EISCAT observations show eastward mean wind with amplitudes of ~25 m s^{-1} at 95 and 98 km and then decreasing with increasing height. Both altitude profiles appear to be reasonable, but a large difference (~15 ms^{-1}) between EISCAT and MF radar winds is found at 100 km height. The EISCAT wind data are averaged over 17 days, obtained from a period 1987-1996, while the MF radar wind are monthly means in January and December 1999, when the solar activity was high. These facts suggest that winter zonal mean wind may be more variable with month, year and/or solar activity conditions than that in summer.

The semidiurnal amplitudes derived by the MF radar in winter months are about 10-20 m s^{-1} between 90 and 100 km. This feature is similar to that of the earlier study by *Manson et al.* [1999], who compared tidal components derived from the MF wind data with those predicted by Global Scale Wave Model (GSWM) [see e.g., *Hagan et al.*, 1999]. In their comparison, they showed a general tendency at all locations for observed values in winter above 85-90 km to be smaller than the model, while in all other months the MF wind amplitudes were larger than GSWM-95 often considerably. Bearing in mind the 1.5 amplitude factor found by *Manson et al.* [1992] for the MF wind amplitude, and applying that to these profiles, the semidiurnal amplitudes become much stronger. This plus the excellent phase agreements, suggest that tidal MF radar 12-h data near 100 km for winter are a valuable product. In contrast, concerning the diurnal tide, the amplitude and phase from the 2 radars are not in agreement for winter. There are some large amplitudes from EISCAT below 100 km. The amplitude profiles of the MF radar wind is again similar to that of *Manson et al.* [1999] who showed a relatively good agreement (the modeled amplitude laying within the range of observed values over much of the height range) of the diurnal amplitude between MF radar wind observations and GSWM-95 predictions for all months. If we exclude these large values from the EISCAT radar wind profiles and combine the two altitude profiles at ~97-100 km from the different radar observations, the altitude profile is reasonably well behaved from 70 to 116 km. Therefore those large amplitudes from the EISCAT radar might be wrong, but this should be left for future studies to make final judgment.

Electric fields are also shown in Figures 1a, 2a, 2b, and 3a to examine a possible relationship with wind velocity obtained by the MF radar. The MF radar wind measurements depend upon scatter from electron density irregularities associated with turbulence or sharply bounded layers [*Thrane et al.*, 1987]. However, in the lower thermosphere and above 95 km, there is a tendency for electrons to decouple from the neutrals, a process countered by ambi-polar diffusion within various turbulence and irregularity structures [*Käiser*, 1969; *Reid*, 1983]. In the latter paper, the decoupling process was shown to be enhanced by E-fields associated with auroral activity. It appears, however, that no such effect can be identified between the E-field and the MF radar wind data in these events we have examined. Thus, it is probable that up to ~100 km, the MF radar observes the neutral wind proper without a contamination of the $\mathbf{E}\times\mathbf{B}$ drift velocity of electrons.

Finally Figure 5 shows seasonal variations of the horizontal mean winds. The wind data obtained by the MF radar in 1999 and the EISCAT radar in 1987-1996 are shown for the height range from 70 to 91 km and from 95 to 119 km, respectively. Although the wind data were obtained at different times under different solar activity conditions, they show clear variations in season and altitude from the mesosphere to the lower thermosphere. For the zonal mean wind, strong westward flow is seen in the summer mesosphere, while strong eastward flow is found in the lower thermosphere below 110 km in summer. On the other hand, the meridional mean wind in summer blows southward in the mesosphere and northward in the lower thermosphere. It is interesting to notice that the amplitude of the meridional mean wind in summer maximizes at ~10 km higher than that of the zonal mean wind both in the mesosphere and the lower thermosphere in agreement with zonal and meridional wind contour plots from a range of radars (35°S - 70°N) shown by *Manson et al.* [1991]. The Coriolis torque on the summer southward flow (acting westward) was therefore shown to maximize in the region of maximum eastward shear; it was argued that during a time of relatively constant mean zonal wind the gravity wave drag would be equal and opposite (eastward). Similar relationships are therefore appropriate here. Figure 5 implies that our understandings of thermospheric dynamics at high latitude will be significantly improved by combining wind data from the EISCAT and MF radars.

6 Summary and conclusions

Using wind data obtained by the EISCAT UHF radar and MF radar co-located in Tromsø, Norway, we have made a comparison study of winds obtained both as case studies and based on statistical treatment. For more than 20 days from February to October in 1999, simultaneous wind data obtained by EISCAT and MF radars are compared in terms of temporal wind variations. Based on electron density profiles obtained simultaneously by the EISCAT radar, the difference between real and virtual heights caused by the effect of group retardation for 2.8 MHz radio wave is calculated. Solar EUV effect on the group retardation is likely significant above 90 km in summer, but at and above 105 km in winter. It is shown that at a high latitude station like Tromsø an effect of particle precipitation is important and affects wind observations based on the MF radar. Sometimes, the effect is seen at 84 km or lower.

We have also made comparisons of mean, diurnal and semidiurnal winds. From 1-month-averaged wind data obtained by the Tromsø MF radar, we have derived mean, diurnal and semidiurnal winds for each month in 1999. These are compared with the statistical wind profiles obtained by *Nozawa and Brekke* [1999a] that were based on 56 days of the EISCAT wind data obtained from 1987 to 1996 under geomagnetically quiet conditions ($Ap < 16$). From the statistical comparison, we showed a good possibility for combining wind profiles from the two radars. Finally, it should be noticed that care is necessary for MF radar wind data above 91 km for summer obtained at a high latitude station. In the height range 90-100 km for winter and equinoxes, the two systems appear complementary. There is some variability with height for both systems in winter in the region, which is probably associated with reduced ionization and the relatively decreased signal to noise ratio for the scatter from both radars.

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Figure captions

Figure 1a. Upper two panels electric field of meridional (positively north) and zonal (positively east) components are shown from 10 UT on February 9 to 16 UT on February 12, 1999 obtained by the EISCAT CP-1 mode with error bars (1σ). Next four panels wind data obtained by EISCAT (pluses) and MF (filled circles) radars are illustrated as a function of time for four heights between 90 and 99 km. The vertical bars represent averaged errors for EISCAT velocity values. At the bottom, 3-h Kp index is also shown over the experiment period.

Figure 1b. The difference in km between virtual and real heights for 2.8 MHz radio wave are shown as a function of UT for 8 heights from 84 to 99 km from 10 UT on February 9 and 16 UT on February 12, 1999. The difference is calculated by using altitude profiles of electron density obtained by the EISCAT radar (see text). Solid circles with a value of -1 mean that total reflection occurs below the height, and solid circle with a value of 9 means that the difference is greater than 9 km. Vertical bar associated with each value represents possible ambiguity when electron density is higher and lower by 10% . At the bottom, height integrated Hall conductivities (conductances) are shown over the experiment period.

Figure 2a. Upper two panels, northward and eastward components of electric field obtained by EISCAT CP-2 mode from 15 UT on July 1 to 16 UT on July 9 are shown. Next four panels, meridional wind velocity (positively northwards) obtained by EISCAT (pluses) and MF (dots) radars are shown for 4 heights from 89 km to 98 km. Bottom panel: 3-h Kp index is also shown.

Figure 2b. Same as Figure 2a except for zonal wind data (positively eastwards).

Figure 2c. Same as Figure 1b except for the period from 15 UT on July 1 to 16 UT on July 9, 1999.

Figure 3a. Same as Figure 1a except for the period from 15 UT on September 15 to 16 UT on September 17, 1999.

Figure 3b. Same as Figure 1b except for the period from 15 UT on September 15 to 16 UT on September 17, 1999.

Figure 4a. Mean winds derived by the MF radar and EISCAT radar observations (solid circle) are compared. The values of the MF radar are based on 1-month averaged wind data obtained in 1999, while the EISCAT wind data are based on 56 days obtained from 1987 to 1996. Since EISCAT data are sorted to 3 seasons, wind data by MF radar obtained in June and July, March, April, September and October, and January and December are compared with EISCAT wind data for summer, equinox and winter, respectively. Following work of *Nozawa and Brekke* [1995] to estimate ambiguities of derived values, a standard deviation (1σ) associated with each data value is also shown.

Figure 4b. Same as Figure 4a except for diurnal amplitudes and phases.

Figure 4c. Same as Figure 4a except for semidiurnal amplitudes and phases.

Figure 5. Seasonal variation of meridional and zonal mean winds are shown at upper and lower panels respectively from 70 km to 119 km. From 70 km to 91 km, MF radar wind data are used and at and above 95 km EISCAT radar wind data are shown. The MF radar wind data were obtained over 1999 and averaged for 1-month, while the EISCAT wind data were obtained from 1987 to 1996 and were divided onto 3 seasons.

Table 1 List of EISCAT datasets.

Dataset	Start		End		Observation period(hr)	Mode	Averaged T_{sys}[1] (K)	Averaged T_x[2] power (kW)
	date	time (hr)	date	time (hr)				
990209	Feb 9	10	Feb 12	16	78	CP-1	96	1122
990308	Mar 8	10	Mar 12	12	98	CP-2	102	1216
990701	July 1	15	July 9	16	193	CP-2[3]	156	1086
990915	Sep 15	15	Sep 17	16	49	CP-1	108	1092
991012	Oct 12	10	Oct 15	16	78	CP-1	99	1143

[1] T_{sys} stands for the system noise temperature.
[2] T_x stands for transmitter.
[3] Conducted as a Special Program by the seven EISCAT associated countries for almost half of time.

Table 2 Geomagnetic activity.

Date	3-h Kp index in UT								Ap	$F_{10.7}$
	0-3	3-6	6-9	9-12	12-15	15-18	18-21	21-24	index	index
990209	2-	1+	1-	0	0+	1-	1-	1-	3	125.9
990210	1	1	2	0+	1-	2-	2+	2+	6	148.4
990211	4-	3-	3	3	4	4	4	3	20	159.3
990212	4-	4-	4+	4	3+	4+	4	2+	24	183.6
990308	3-	4-	2-	2-	1+	2	2+	4+	12	125.0
990309	4-	4	4	3+	4+	3+	3+	2-	21	125.3
990310	5	6+	5-	5	2+	3-	3-	2+	34	133.6
990311	3+	3+	3+	3	4-	3	2	2-	15	135.3
990312	1+	2+	2+	3	3-	2+	3+	3-	11	138.5
990701	1	1-	1+	1+	1+	1	3-	3-	6	202.0
990702	4-	3+	5	4	2+	4	4	4+	26	193.4
990703	3	2+	4-	2-	1-	2	0+	1+	9	203.5
990704	0	0	0+	0+	1-	0+	0+	1-	2	191.9
990705	0	0	0	0+	1-	1	1+	2-	3	180.0
990706	3-	1+	2-	1-	1-	3	3-	2+	8	173.5
990707	1+	1+	1-	0+	0+	1	2	2-	4	163.7
990708	2	1	1	2-	2+	2-	2-	2	6	154.1
990709	2+	1	1-	1-	1-	0+	1-	1-	4	155.9
990915	4-	4-	5	6-	4-	3-	3	2-	27	156.6
990916	3+	6	5	4+	3+	4-	3	3-	31	159.9
990917	3	2+	2	2	4	4+	4-	2+	16	159.1
991012	5	5	4-	5	5+	5+	5+	4+	46	182.9
991013	5-	3	4	3-	4-	4	4	3+	23	190.1
991014	3+	4+	4-	4-	5-	5	4+	4+	31	198.7
991015	5	5	4+	4	4-	5-	3+	3-	31	197.1