Long-term observations of polar mesosphere summer echoes using the ESRAD MST radar

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Abstract

Polar Mesosphere Summer Echoes (PMSE) are strong radar echoes observed from altitudes of 80-90 km in polar regions, during summer time. PMSE are closely related to the fascinating atmospheric phenomenon known as noctilucent clouds (NLC). Since it has been suspected that NLC could respond to climate change in the mesosphere, they have attracted considerable interest in the scientific community during recent years. However, continuous visual or photographic NLC observations suffer from weather restrictions and the human factor. In contrast, PMSE radar measurements can easily be made over a long interval and are very attractive for long-term studies of the atmospheric parameters at the polar mesopause.

This thesis uses the world’s longest data set of PMSE observations made by the same radar at the same place. Since 1997 these measurements have been carried out with the 52 MHz ESRAD MST radar located near Kiruna in Northern Sweden. The data set for 1997-2008 has been used for studies of diurnal, day-to-day and year-to-year variations of PMSE. We showed that PMSE occurrence rate and volume reflectivity on a daily scale show predominantly semidiurnal variations with the shape of the diurnal curves remaining consistent from year to year. We found that day-to-day and inter-annual variations of PMSE correlate with geomagnetic activity while they do not correlate with mesopause temperature or solar activity. We did not find any statistically significant trends in PMSE occurrence rate and length of PMSE season over 1997-2008.

The thesis also presents a new, independent calibration method, which can be used to estimate changes in transmitter output and antenna feed losses from year to year (for example due to changes of antenna configuration) and allows making accurate calculations of PMSE strength. This method is based on radar-radiosonde comparisons in the upper troposphere/lower stratosphere region simultaneously with PMSE observations. Using this calibration we calculated the distribution of PMSE strength over magnitudes; it varies from year to year with the peak of the distribution varying from $2\times10^{-15}$ to $3\times10^{-14}$ m$^{-1}$. We found that inter-annual variations of PMSE volume reflectivity strongly correlate with the local geomagnetic k-index and anticorrelate with solar 10.7 cm flux. We did not identify any significant trend in PMSE volume reflectivity over 1997–2009.

Finally, using 11 years of measurements, we calculated in-beam the PMSE aspect sensitivities using the FCA technique. We showed that half of PMSE detected each year cannot be explained by isotropic turbulence since they are highly aspect sensitive echoes. The distribution of these echoes remains consistent from year to year with median values of aspect sensitivity from 2.9 to 3.7°. The remaining half of the PMSE have aspect sensitivity parameters larger than 9-11°. We found that PMSE aspect sensitivity has altitude dependence: the scatter becomes more isotropic with increasing height. We did not identify any dependence of PMSE aspect sensitivity on backscattered power for any year. We analysed limitations of the in-beam and off-zenith beam methods and concluded that the former is suitable for highly aspect sensitive echoes while the latter is needed for more isotropic scatterers.
Sammanfattning

Polarmesosfäriska sommarekon (PMSE) är ett fenomen där starka radarekon genereras vid altituder av 80-90 km i polarregioner under sommartid. PMSE är nära besläktat med det fascinerande atmosfärssfenomen som är känt som naturligt moln (NLC). På grund av misstankar om att NLC kan reagera på klimatförändringar i mesosfären har det attraherat ett betydande intresse från det vetenskapliga samfundet på senare tid. Kontinuerliga, visuella eller fotografiska NLC-observationer lider dock av väderbegränsningar och den mänskliga faktorn. Radarmätningar av PMSE, däremot, kan lätt utföras över ett längre tidsintervall och är mycket attraktiva för längtidsstudier av de atmosfäriska parametrarna vid den polära mesopausen.


I denna avhandling presenterar vi även en ny, oberoende kalibreringsmetod, vilken kan användas för att uppskatta förändringar i sändareffekt och antennmätningstödsförluster från år till år (till exempel på grund av förändringar av antennkonfiguration) och tillåter träfshårliga beräkningar av PMSE-styrka. Denna metod baseras på radar-radiosondjämförelser i övre troposfären/lägre stratosfären, tillsammans med PMSE-observationer. Med denna kalibreringsmetod beräknade vi PMSE-styrkofördelningen, vilken hade årliga toppar som varierade från $2 \times 10^{-15}$ till $3 \times 10^{-14} \text{ m}^{-1}$. Vi fann att år till år-variationer av PMSE-volymreflektivitet starkt korrelerar med det lokala geomagnetiska k-indexet och anti-korrelerar med flödet av solär 10.7 cm-radiostrålning. Vi identifierade ingen signifikant trend i PMSE-volymreflektivitet över 1997-2009.

List of included papers

This thesis is based on the following papers, which are referred to in the text by their Roman numerals.


Other papers by the author

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1. Introduction

Since the 1980s it has been known that during summer time, extremely strong radar echoes can be detected from the mesopause altitudes at polar regions. These echoes are known as Polar Mesosphere Summer Echoes (PMSE) and are closely related to the visible phenomenon known as noctilucent clouds (NLC). PMSE are linked to important atmospheric parameters and processes such as mesospheric temperature, planetary and gravity waves, winds, turbulence and can be used for continuous monitoring of the thermal and dynamic state of the mesopause region. In the past, many observations of mesospheric echoes have been made using radar, lidar, rockets and satellites. However, some of the observed PMSE features are not yet fully studied.

The studies of long-term trends and variations in the polar mesosphere play a special role in atmospheric research. In recent years, these investigations attracted considerable scientific interest since it has been suggested that noctilucent clouds observed in the polar mesosphere could follow the global climate change (Thomas and Olivero, 2001). However, it is difficult to test this hypothesis since the long-term observations of NLC are based mainly on visual and lidar observations which depend on human and weather factors. Satellite measurements have many advantages, but have relatively small observation periods for long-term investigations. For example, the AIM satellite mission that was launched in 2007 was dedicated to exploration of NLC. Thus PMSE radar measurements over a long interval, at least one solar cycle, provide unique information for studying variations and long-term trends in atmospheric parameters since they are independent of observers and weather conditions.

The majority of PMSE observations have been made using mesosphere-stratosphere-troposphere (MST) radars, operating in the lower VHF band, usually near 50 MHz. These radars are used to investigate the dynamics of the middle and lower atmosphere and provide invaluable information about turbulence, wind, waves etc. This thesis is based on PMSE measurements carried out since 1997 with the 52 MHz ESRAD MST radar, located near Kiruna, in northern Sweden. It is the world’s longest data set of PMSE observations made by the same radar at the same place. The papers included in this thesis present the studies of different properties of PMSE observed over a long time interval as well as a new method of radar calibration.
2 Polar Mesosphere Summer Echoes

2.1 Vertical structure of the Earth atmosphere

The Earth atmosphere is divided into layers according to the vertical structure of the temperature field. It consists of four layers: the troposphere, the stratosphere, the mesosphere and the thermosphere, which are separated by the tropopause, the stratopause, and the mesopause (Fig.1).

![Diagram of Earth's atmosphere layers](image)

Fig.1. Thermal structure of the Earth atmosphere. (After Brasseur and Solomon, 1986)

The troposphere extends from the surface to about 8 km at the pole (18 km at the equator). In the troposphere the temperature is decreasing with altitude, due to the absorption and re-radiation of incoming solar energy at the surface. The boundary between the troposphere and stratosphere is known as the tropopause. The tropopause is characterised by changing of the environmental lapse rate from positive (in the troposphere) to negative (in the stratosphere). The stratosphere extends from the tropopause region to the height of about 50 km. Temperature in the stratosphere increases with altitude, due to increased absorption of solar ultraviolet
radiation by the ozone layer. The stratopause, a boundary between the stratosphere and mesosphere, is the area where a maximum in the temperature occurs, typically at 50 - 55 km. The mesosphere, which is characterised by decrease of temperature, extends from 55 km up to 90 km. In this thesis we focus on the polar mesosphere. The top of the polar mesosphere, the mesopause, is the coldest part of the Earth's atmosphere, with temperatures falling down to 130 K. A number of interesting phenomena take place in the polar mesosphere. Most meteors typically occur and burn up there. They can be observed by the naked eye between about 65 and 120 km above the Earth. The most fascinating visual phenomenon in the mesosphere is noctilucent clouds which appear in the polar regions in both hemispheres during summer. However, the mesosphere is hard to access for study. Balloons and jet planes cannot fly high enough to reach it; only rockets can reach this region, providing infrequent and expensive in-situ measurements. In this context, continuous remote sensing by satellites and by ground-based instruments such as radars plays an essential role in mesospheric studies.
2.2 Dynamics and thermal state of the polar mesosphere

The globally-averaged temperature field at individual altitudes in the middle atmosphere is in approximate radiative equilibrium. However, especially near the polar mesopause, eddy motions (waves) cause significant local departures from equilibrium. As a result, the summer polar mesosphere is much colder than radiative equilibrium and the winter mesosphere much warmer.

In the atmosphere, the zonal wind is in balance with the latitudinal gradient of temperature, as described by the thermal wind equation

\[
\frac{\partial u}{\partial z} = -\frac{r}{fHR} \frac{\partial T}{\partial \phi},
\]

where \( u \) is the zonal wind velocity, \( z \) is the vertical coordinate, \( f \) is the Coriolis parameter, \( r \) is the gas constant for air, \( R \) is the radius of the Earth, \( H \) is the scale height, \( T \) is the temperature and \( \phi \) is the latitude.

Fig.2. Schematic latitude-height section of typical observed zonal mean wind speeds; contours in m/s. W and E denote centers of westerly (from the west) and easterly (from the east) winds, respectively. Negative values denote winds from the east. (After D. G. Andrews, J. R. Holton, and C. B. Leovy, Middle Atmosphere Dynamics, Academic Press, 1987)

The highest temperature in the lower troposphere is located at the equator and it decreases towards the poles. This distribution of temperature gradient leads to increasing eastward zonal wind with height. In the stratosphere, temperature reaches
the maximum values in the summer hemisphere. This leads to dominating westward winds in the summer stratosphere and eastward winds in the winter stratosphere. The schematic latitude-height cross section of the solstice mean zonal wind fields, averaged over longitudes, is shown in Fig. 2. Gravity waves propagate from sources in the troposphere (like convection motion, flow over mountains, etc.) to the upper part of atmosphere, interacting with zonal flow. This causes the gravity waves with eastward propagating mode to reach the summer mesosphere, where they break. Conversely, in winter, the westward propagating gravity waves reach the mesospheric heights. These gravity waves break and deposit momentum flux and energy to the mesosphere that decelerates the zonal wind and, by thermal wind balances, changes the temperature field. This gives rise to meridional circulation, which is characterized by upwelling of the air mass at the summer pole and downwelling at the winter pole, with zonal-mean meridional flow from the summer to the winter hemisphere at mesopause heights (Fig 3).

Fig. 3. Schematic streamlines of the solstice diabatic circulation in the middle atmosphere. (After D. G. Andrews, J. R. Holton, and C. B. Leovy, Middle Atmosphere Dynamics, Academic Press, 1987)

When the air mass is lifted upwards from its equilibrium level, the volume is expanding adiabatically on the expense of internal energy of the parcel, which leads to temperature decrease. Thus during summer time in the northern hemisphere, an equatorward wind is associated with the transport of cold air and a poleward wind with the transport of warm air. This circulation explains why the polar mesopause is
much colder during summer than winter. The cross section of the mean temperature, averaged over longitudes, for solstice conditions is shown in Fig.4.

Fig.4. Schematic latitude-height section of zonal mean temperature (°C) for solstice conditions. (After D. G. Andrews, J. R. Holton, and C. B. Leovy, Middle Atmosphere Dynamics, Academic Press, 1987)
2.3 Noctilucent clouds and PMSE

As a result of meridional circulation induced by gravity waves, temperatures of ~130 K and lower are reached from May to August at mesopause altitudes in the Northern hemisphere (Lübken et al., 1999). These low temperatures lead to nucleation and growth of ice particles from water vapor at around ~85-88 km. The radii of these particles were found to be between ~5 and 70 nm (Hervig et al., 2009). The largest ice particles (larger than ~20 nm) can be seen from the ground, by the naked eye, as a silver-grey pattern known as noctilucent clouds, NLC (Fig. 5). The first observation of NLC is dated to 1883 (Leslie, 1885). One century later, it was found that this phenomenon of noctilucent clouds is accompanied by strong radar echoes detected by VHF radars (Ecklund and Balsley, 1981). These echoes are known as Polar Mesosphere Summer Echoes (PMSE).

![Fig.5. Noctilucent Clouds over Moscow, 27 June 2005 (Photo by P. Dalin).](image)

The common volume observations of NLC and PMSE showed that an NLC layer takes place at the lower edge of the radar echo layer (Nussbaumer et al., 1996, von Zahn and Bremen, 1999). All these observations are in agreement with the growth and sedimentation scenario for ice particles in the mesosphere. In this scenario, ice particles nucleate close to the mesopause level, where the levels with highest supersaturation can be found. Their growth continues by condensation of water vapor while they sediment down due to gravity. Finally they become large enough to be visible as NLC (Rapp and Lübken, 2004).
The radar echoes occur due to scattering from fluctuations in the electron density. Changes in the electron density can be caused by several factors. In order for PMSE to occur, a combination of atmospheric turbulence produced by gravity waves and electrically charged ice particles is needed. PMSE occur at heights of 80 – 90 km, where ice particles are formed from the water vapor under the very low temperature conditions. These particles were measured in several rocket experiments (Havnes et al, 1996, 2001; Mitchell et al., 2003, Smiley et al., 2003, Strelnikov et al, 2009). They are immersed in the D-region plasma, where free electrons stick to their surfaces, resulting in charged particles. In addition, gravity waves break at the same altitudes, producing turbulence. Turbulence causes uneven distribution of ice-charged particles and by that also irregular spatial distribution of electron number density, which leads to irregularities in the refractive index observed by VHF radars as PMSE. Fig. 6 shows a typical example of PMSE observed by the ESRAD radar during one of the day in 2010.

Fig.6. Backscattered power measured with ESRAD MST radar on, 9 July 2010 over Kiruna. Colour scale indicates log10 of volume reflectivity $g$. 

2.4 PMSE characteristics

2.4.1 Occurrence rate and volume reflectivity

In general, PMSE is considered to occur if the echo strength exceeds a certain threshold during a given time interval within a given altitude range. The majority of previous PMSE studies were based on various PMSE detection threshold values expressed in signal-to-noise ratio (SNR). For example, Barabash et al. (1998) set SNR$_{th}$=-5 dB for the ESRAD radar and Bremer et al. (2009) used SNR$_{th}$=4 dB for the ALWIN radar. The occurrence rate (OR) calculated using PMSE detection threshold in terms of SNR makes them dependent on radar parameters as well as PMSE strength. Even the usage of the same threshold, for radars with different noise levels, implies different signals, i.e. PMSE strengths. Therefore, OR calculated according to this procedure is not suitable for comparative studies of PMSE.

Another characteristic of PMSE, the radar volume reflectivity $\eta$ (or radar scattering cross section), describes PMSE absolute strength and can be used for comparison of PMSE measured with different radars at different locations. $\eta$ is defined as the power which would be scattered isotropically with a power density equal to that of the backscattered radiation, per unit volume and per unit incident power density, and can be calculated according to the formula (e.g. Gage, 1990):

$$\eta = \frac{P_t}{P_r} \frac{64(2\ln 2)r^2}{\pi LAV_f \Delta r}$$

(2)

where $P_t$ is the power delivered to the radar, $P_r$ is the power received by the radar, $r$ is the distance to the scattering volume, $\Delta r$ is the range resolution along the radar beam, L is the loss in the antenna feed, $V_f$ is the fraction of the scattering volume which is filled with scatterers ($V_f=1$ was usually assumed) and $A_e$ is the effective area of the receiving antenna.

In our investigation we calculated OR based on threshold of volume reflectivity, i.e. OR was equal to 1 if $\eta$ exceeds a value of $10^{-15}$ m$^{-1}$ at any altitude in the 80-90 km range.
2.4.2 Short-term variations

PMSE OR and volume reflectivity show a strong variability on different time scales. One can classify these variations on the basis of the length of the time interval in which they occur: short-term and long-term. As short-term variations, we consider diurnal and day-to-day (or seasonal) variations. Long-term are year-to-year variations.

The diurnal variations of PMSE have been observed by several radar campaigns (Czechowsky et al., 1989; Williams et al., 1995; Kirkwood et al., 1995; Hoffmann et al., 1999). Hoffmann et al. (1999), based on data collected during four PMSE seasons between 1994-1997 at Andenes, Norway, derived the mean diurnal variation of the occurrence of PMSE as a function of local time and height. They found that PMSE on a daily scale shows mainly semidiurnal variation with maxima at noon (12-13 LT) and around midnight and minima near 6-7 LT and 18-19 LT. Similar semidiurnal variations of PMSE OR were observed also in the Southern Hemisphere, at Davis, Antarctica, for seasons 2004-2005, by Morris et al. (2007). We consider the diurnal variations of both PMSE OR and volume reflectivity and their possible changes from year to year using PMSE measurements for 1997-2008 in Paper I. In spite of many observations of PMSE variability on a daily scale, the mechanism behind them is not yet completely clear. Hoffmann et al. (1999) found that this behaviour of daily curves is markedly influenced by temperature changes induced by tidal waves in the mesospheric wind field. Klostermeyer (1999) proposed a more complex explanation for diurnal variation of volume reflectivity, which includes tidal temperature variation, electron production by solar ionization and by energetic particle precipitation.

The seasonal variation of PMSE is known since its first observation, carried out at Poker Flat, Alaska (Ecklund and Balsley, 1981). The recent long-term observations of PMSE at Andenes during the years 1999-2008 showed that PMSE in the Northern Hemisphere usually start near 20 May, increase sharply until the beginning of June, stay at the highest level until the end of July and then decrease gradually until the end of August (Bremer et al., 2009). This seasonal PMSE behaviour is mainly explained by the seasonal variations of the mesospheric water vapor content and temperature (Bremer et al., 2003). The latter explanation is supported by e.g. measurements of OH temperature over Stockholm by Espy and Stegman (2002).
They noticed similarity in asymmetry of the shape of summer temperature minimum and PMSE occurrence rate.

PMSE occurrence rates were observed to be strongly influenced by planetary waves (Kirkwood and Rechou, 1998, Kirkwood et al., 2002, Klekociuk et al., 2008). Mesospheric temperature fluctuations due to these waves were found to be a primary factor affecting PMSE OR in the beginning and the end of the season (Kirkwood et al., 2002). Morris et al. (2007, 2009) found the modulation of the PMSE OR envelope at Davis, Antarctica by mesopause temperature variations caused by changes in meridional wind. However, this modulation may rather be associated with 5-day planetary waves than with other factors.

Zeller et al., (2009) have investigated a possible influence of mesospheric temperature on PMSE SNR with ALWIN radar at Andenes for two PMSE seasons. They reported that at the start and end of the PMSE season, the correlation is significant and strongly negative, but in the middle of the season, the correlation is weaker and not significant. The authors concluded that temperature variations have only a low impact on PMSE variations when the temperature is significantly below the frost point.

Zeller and Bremer (2009) using measurements from the ALWIN MST radar in Andenes for the years 1999-2005 reported that PMSE OR was stronger on 1-2 days after the maximum Ap index for moderate geomagnetic activity. However, the significance level of this result was relatively low. With higher significance the authors showed that PMSE OR decreases on the day when local K index reaches its minimum.

In Paper I we aim to find which factors other than planetary waves can influence PMSE behaviour from day to day. We examined the possible relationship between day-to-day variations of PMSE OR, mesopause temperature, meridional wind and geomagnetic activity.
2.4.3 Long-term variations and trends

In order to accurately assess possible secular trends in strength and occurrence of PMSE, one needs firstly long data series covering at least one solar cycle and secondly elimination of any variations related to this solar cycle from the data.

The first year-to-year PMSE observations over an interval as long as a solar cycle were reported by Bremer et al. in 2003, and later updated by the same authors in 2006 and 2009. These measurements were carried out at Andenes, Norway (69.3°N; 16.0°E), in 1994-2008 with two different VHF 53.5 MHz radars: ALOMAR SOUSY radar (1994-1997) and ALWIN radar (1999-2008). The possible long-term variations of PMSE were investigated in terms of their occurrence rate and seasonal duration. The value of the signal-to-noise ratio (SNR) was used as the detection threshold of PMSE which was different for different radars/years. The authors examined the dependence of PMSE on solar and geomagnetic activity changes in a solar cycle scale. The results are shown in Fig.7.

![Fig.7. Dependence of the PMSE occurrence rate and solar Lyman α radiation (upper panel) and on Ap index (lower panel) (After Bremer et al., 2009)](image-url)
In both cases, a positive correlation was found. The correlation of PMSE OR with Ap is significant to a level of more than 99%. The correlation with Lyman $\alpha$ is much less significant (<90%). In order to derive a trend, the authors used the twofold regression analysis for removing variations induced by solar and geomagnetic activity which could have influence on variation of PMSE. The long-term variation of the PMSE occurrence rate after elimination of their solar and geomagnetic caused parts is presented in Fig. 8.

![Fig.8](image)

**Fig.8.** Long-term variation of the PMSE occurrence rate after elimination of their solar and geomagnetic caused parts (After Bremer et al., 2009)

The authors also investigated the long-term variations of PMSE in terms of their seasonal duration. Finally, the authors reported that positive trends of 0.41%/year and 0.37 days/year could be detected in the PMSE OR and the duration of PMSE season, correspondingly. However, due to relatively short data series, the statistical significance level of these trends is relatively small.

In paper I and II we have investigated the influence of solar and geomagnetic activity on year-to-year variations in PMSE, expressed in terms of occurrence rate, volume reflectivity and length of the season. Statistical evaluations of trends have been made for PMSE OR, $\eta$ and seasonal duration.
### 2.5 Aspect sensitivity

Many PMSE studies reveal that the echoes are highly aspect sensitive. Aspect sensitivity is a property of the scatterers which describes the variation of scattered power in respect to incident angle. A backscatter angular polar diagram of the scatterers $P(\theta)$ can be approximated as (Hocking, 1986):

$$P(\theta) = \exp\left[ -\frac{\sin^2 \theta}{\sin^2 \theta_{as}} \right],$$  \hspace{1cm} (3)

where $\theta$ is the zenith angle and $\theta_{as}$ is $e^{-1}$ half-width of the polar diagram, which is called the aspect sensitivity parameter. A polar diagram with $\theta_{as}=20^\circ$ is shown in Fig. 9.

Aspect sensitivity of radar echoes can provide us information about the scattering process and the nature of scatterers. Two ‘extreme’ scattering mechanisms relevant to MST radars are turbulent scatter and Fresnel (partial) reflection. The turbulent scatter is caused by spatial variations of the radar’s Bragg scales due to turbulent processes and is therefore isotropic. Fresnel reflection arises from horizontally stratified stable layers which have a horizontal scale of at least one Fresnel zone and is highly aspect sensitive. However, it later became obvious that these theories cannot explain many observed features of PMSE and the models of Fresnel scatter and anisotropic turbulent scatter need to be involved for explanation of the scattering process in the PMSE layer (e.g., Balsley and Gage, 1980; Röttger, 1980). Fresnel scatter models propose horizontal layers with vertically varying refractive index and can better explain the case when the power drops very rapidly as
a function of zenith angle. The anisotropic turbulence model is more useful for explaining the situation when backscattered power changes uniformly up to 30° of the off-zenith angle.

The measurement of aspect-angle sensitivities can be accomplished by two methods: (1) by comparing the echo strengths from vertical and off-vertical radar beams and (2) by in-beam estimates using spaced antennas (SA). The latter applies coherent radar images (CRI) or the full correlation analysis (FCA) techniques. In the majority of PMSE aspect sensitivities studies, the radar-tilted beam was used (Czechowsky et al., 1988; Hoppe et al., 1990; Huaman and Balsley, 1998; Swarnalingam et al., 2011). The studies found θ_{as} to be mostly in the range 8°-13°. When the in-beam method was used (Zecha et al., 2001, Chilson et al., 2002, Paper III), it gave smaller values for θ_{as} of 1-5°, i.e. PMSE were highly aspect sensitive. Both methods have some disadvantages and limitations. We discussed limitations in relation to the finite radar beam in Paper III. Some studies (e.g. Czechowsky et al., 1988, Chilson et al., 2002, Swarnalingam et al., 2011) reported the dependence of PMSE aspect sensitivity on altitude, e.g., that lower PMSE layers are associated with more anisotropic scattering.

There is one more important point of PMSE aspect sensitivity: If PMSE are strongly aspect sensitive, this can affect estimation of their absolute strength, i.e. volume reflectivity, because echoes do not fill the entire radar sampling volume. Swarnalingam et al. (2009a) found that for the Resolute Bay VHF radar with a narrow beam of 1.4° (two-way) width and PMSE with θ_{as}=5°, this effect is not significant. However, in Paper III we showed that for the ESRAD radar, the effect of PMSE high aspect sensitivity can result in an underestimation of echo volume reflectivity by a factor 2.
3 MST radars for PMSE observations

3.1 Radar scattering cross section

As mentioned in Chapter 2, MST radars have several advantages over satellite and in-situ measurements, and are invaluable for long-term studies of the mesospheric region.

The radar echoes measured by MST radars are caused by scattering from irregularities in the atmospheric refractive index $n$, revealing structures at the radar half wavelength (Tatarskii, 1961). This scatter is called Bragg scatter. The variations in refractive index are related to atmospheric parameters such as electron density, temperature, pressure, and humidity. The refractive index for MST region is defined as:

$$n = 1 + \left( \frac{77.6 P}{T} + 3.75 \cdot 10^4 \frac{e}{T^2} \right) \cdot 10^{-6} - 40.3 \frac{n_e}{f_0},$$

where $P$ is the atmospheric pressure, $T$ is the absolute temperature, $e$ is the partial pressure of water vapour, $n_e$ is the electron number density, and $f_0$ is the radar operating frequency (Balsley and Gage, 1980). The term, which is proportional to pressure, is called the dry term and is dominant for the refractive index in the upper troposphere and stratosphere. The term, which is proportional to humidity, is known as the wet term and plays the main role in the middle troposphere. The influence of these terms can be neglected for the mesosphere, where the refractive index is mainly determined by the electron density. Radar volume reflectivity is determined by fluctuations in the refractive index, i.e. for the PMSE case by those in electron density.

Recently, an expression for PMSE volume reflectivity was proposed by Varney et al. (2010). It includes electron density, turbulence parameters, as well as parameters of charged dust (or ice particles) such as their number density, charge and size. The formula reads as the following:

$$\eta(k) = 8\pi^2 c^2 \frac{f_d q R_i}{\text{Pr} \alpha_d^i} \sqrt{\frac{e\nu_0}{\tau_0}} S^2 k^{-1} \exp \left( - \frac{q(kq)^2}{Sc} \right),$$

$$S = Z_d \left( \frac{N_x}{N_x + Z_i^2 N_i} \right),$$
\[
\hat{M} = \left( \frac{a_0^2 N_i}{g} - \frac{dN_i}{dz} \frac{N_i}{H_n} \right),
\]
\[
Sc = \frac{\nu_a}{D_2}.
\]

where \(k\) is the Bragg scattering wavevector, \(r_e\) is the classical electron radius, \(f_0\) is the proportionality constant \((f_0=2)\), \(q\) is a positive dimensionless constant, \(R_i\) is the Richardson number, \(Pr^t\) is the turbulent Prandtl number, \(\omega_0\) is the Brunt-Vaisala frequency, \(\varepsilon\) is energy dissipation rate, \(\nu_a\) is the kinematic viscosity of air, \(Z_d\) is the dust charge, \(N_e\) is the electron density, \(Z_i\) is the ion charge, \(N_i\) is the ion density, \(N_d\) is the dust density, \(g\) is the gravitational acceleration, \(H_n\) is the neutral scale height, \(S_c\) is the Schmidt number and \(D_2\) is the effective diffusion coefficient.

However, not every PMSE can be described by the proposed model of isotropic turbulent scatter. This model is not applicable to highly aspect-sensitive PMSE. For these cases, Fresnel scatter/reflection is more appropriate (see Chapter 2).
3.2 ESRAD technical description

ESRAD is an MST-class 52 MHz radar located near Kiruna in northern Sweden. It has been in operation since June 1996 (Fig.10). The radar provides information on the dynamic state of the atmosphere such as waves, winds, layering and turbulence in the altitude range of 1 km to 110 km. The hardware for transmitter and receiver systems was constructed by the Atmospheric Radar System Party (ATRAD), Adelaide, Australia. The radar installation and maintenance have been realized by the Swedish Space Corporation at the ESRANGE rocket range. The data analysis is provided by the Swedish Institute of Space Physics. The first technical description was made by Chilson et al. (1999) and was later updated by Kirkwood et al. (2007). The main parameters of the ESRAD radar are shown in Table 1.

![ESRAD radar in summer 2009 (Photo by M. Mihalikova)](image)

The transmitter radar system has 72 solid-state modules of 1 kW each that are grouped into twelve 6 kW power blocks. Thus the resulting peak power reaches 72 kW, with a maximum duty cycle of 5 %. The pulse repetition frequency can vary from 100 Hz to 16 kHz and the pulse length from 1-50 μs corresponding to range resolution between 150 m and 3 km. The radar is capable of using Baker and complimentary codes for pulse coding of transmitted signals.

Initially, the ESRAD antenna contained a 12×12 phased array of 5-element Yagis, but in April 2004, it was extended up to 16×18. The Yagis are placed at 4.04 m (0.7 times the radar wavelength) from each other and have an approximate height
of 6 m. The antenna array consists of 6 sub-arrays and each sub-array is connected to a separate receiver. The radar operates in the spaced antenna (SA) mode (Briggs, 1984). The ESRAD system is supplied with high level software control and its flexible design allows for future expansions.

Table 1. ESRAD Technical Characteristics

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Location</strong></td>
<td></td>
</tr>
<tr>
<td>Geographic coord./height</td>
<td>67.88°N; 21.10°E; 295 m</td>
</tr>
<tr>
<td>Geomag. Latitude/geomag. Midnight</td>
<td>64.82°N; 22:50 LT</td>
</tr>
<tr>
<td><strong>Transmitter system</strong></td>
<td></td>
</tr>
<tr>
<td>Peak power</td>
<td>72 kW</td>
</tr>
<tr>
<td>Pulse repetition frequency</td>
<td>100 Hz – 16 kHz</td>
</tr>
<tr>
<td>Max duty cycle</td>
<td>5%</td>
</tr>
<tr>
<td>Pulse length</td>
<td>1-50 μs</td>
</tr>
<tr>
<td>Height range</td>
<td>1-110 km</td>
</tr>
<tr>
<td>Codes</td>
<td>Barker (2, 3, 4, 5, 6, 7, 11, 13 bits)</td>
</tr>
<tr>
<td></td>
<td>Complimentary (2, 4, 8, 10, 26, 32, and 64 bits)</td>
</tr>
<tr>
<td><strong>Antenna</strong></td>
<td></td>
</tr>
<tr>
<td>Antenna array</td>
<td>16×18 array 5-element Yagis</td>
</tr>
<tr>
<td>(12×12 array before April 2004)</td>
<td></td>
</tr>
<tr>
<td>Antenna effective area</td>
<td>3740 m² (1870 m² before April 2004)</td>
</tr>
<tr>
<td>Antenna spacing</td>
<td>0.7 λ</td>
</tr>
<tr>
<td><strong>Receiving system</strong></td>
<td></td>
</tr>
<tr>
<td>Sampling interval</td>
<td>1-20 ms</td>
</tr>
<tr>
<td>Filters</td>
<td>250, 500, 1000, and 2000 kHz</td>
</tr>
</tbody>
</table>
3.3 Experiment description

The radar is capable of performing measurements in many different modes. The main radar modes which were used for PMSE long-term studies are ‘fca_4500’ and ‘fca_150’. For investigation of radar-aspect sensitivity we have used the mode ‘fca_150’ which provides the data with the highest available altitude resolution of 150 m. However, since this mode has relatively poor signal-to-noise characteristics, the variations and long-term trends of PMSE have been studied by using ‘fca_4500’ which is able to detect at least ten times weaker echoes than the mode ‘fca_150’. The ‘fca_4500’ mode uses 600 m altitude resolution and a narrower receiver bandwidth which provides the higher sensitivity. The parameters of experiments used in this study are summarised in Table 2.

Table 2. ESRAD operating modes used in this study.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Pulse length (3 dB)</td>
<td>2.1 µs 600 m</td>
<td>1 µs 150 m</td>
<td>1 µs 150 m</td>
</tr>
<tr>
<td>Sampling resolution</td>
<td>1300 Hz 32/128 8 bit</td>
<td>1450 Hz 64/128 8/16 bit</td>
<td>4688 Hz 256 none</td>
</tr>
<tr>
<td>Pulse repetition frequency</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Coherent integrations</td>
<td>Complimentary</td>
<td>Complimentary</td>
<td>none</td>
</tr>
<tr>
<td>Code</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
3.4 Radar Calibration

Radar calibration is necessary for the calculation of PMSE absolute strength expressed in terms of volume reflectivity. It also provides a system-independent comparison of PMSE observed by different radars at different sites. Studies of this kind, addressing PMSE inter-hemispheric and latitudinal differences, have become very popular over the past several years (Kirkwood et al., 2007, Latteck et al., 2008, Morris et al., 2009, Swarnalingam et al., 2009b).

The radar volume reflectivity $\eta$ defined by eq. (2) is proportional to the receiving signal power $P_r$ in W, which can be estimated using the ‘calibration’ equation:

$$P_r = \frac{k_B B \Delta T_{cal} (S_{tot} - (S_{sys} + S_{sky}))}{C_{filt} n_{coh} n_{code} \Delta S_{cal}},$$

(10)

where $k_B$ is Boltzmann’s constant, $B$ is the receiver bandwidth, $\Delta T_{cal}$ is the calibration signal power in K, $S_{tot}$ is the total detected power, $S_{sys}$ is the power due to internal system noise, $S_{sky}$ is the power due to sky noise, $\Delta S_{cal}$ is the calibration signal power, $C_{filt}$ is the receiver filter efficiency, $n_{coh}$ is the number of coherent integrations and $n_{code}$ is a constant that equals to 1 for an uncoded pulse, otherwise twice the number of bits in the code.

For analysis of the ESRAD measurements of PMSE we used the daily variation of galactic noise as the calibration signal for the radar. The sky-map of galactic noise at 53 MHz was calculated by de Olivera-Costa et al. (2008). The authors developed a model that is based on an assimilation of available radio-astronomical surveys and allows for producing a sky-map of any frequency. The daily variation of galactic noise is a convolution of the sky-map with the radar beam pattern, with the radar center moving along the radar latitude on the sky map. The range of this daily variation for the ESRAD is 1680-4500 K. We measured the sky noise daily variation and compared it with the modelled one (Fig.11.).

Note that, in order to confirm that the radar sees the same galactic noise variation as it should according to the radio-astronomical model, ESRAD was also calibrated using noise injection with artificial sources of known strength. The description of this calibration procedure can be found in Kirkwood et al. (2010).
The thesis also presents a new calibration method that, unlike methods previously described, is independent of the radar used. This method allows for estimation of changes in transmitter output and antenna feed losses from year to year (for example due to changes of antenna configuration of ESRAD radar) and can, in principle, be used for improving the accuracy of inter-comparisons of PMSE strength between sites. The antenna loss (L) is the most uncertain parameter for radar volume reflectivity calculations. For the ESRAD radar antenna, maintenance work has been done on several occasions since 1997, which led to corresponding changes in L. We estimated these changes by using a method that is based on radar-radiosonde comparisons in the upper troposphere/lower stratosphere region simultaneously with PMSE observations. A detailed description of this method, together with estimations of effective losses in antenna field in the ESRAD feed for 1997-2010 is presented in Paper II.
4 Summary of the included papers


Paper I reviews PMSE observations carried out with the ESRAD radar during 1997-2008. The characteristics of PMSE were studied in terms of diurnal, day-to-day and year-to-year variations of occurrence rate (OR) and diurnal variations of volume reflectivity (VR). We found that PMSE OR and VR on a daily scale show predominantly semidiurnal variations. The shape of the daily curves is similar for all of the 11 years but minimum and maximum values vary from year to year. We showed that the PMSE season in Kiruna usually starts around May 21-22, reaches its peak by the beginning of June, remains at the highest level until the end of July and then gradually decreases towards the end of August every year. The length of the PMSE season varies from 79 days in 2000 to 99 days in 2003. We found that the start of the PMSE season is associated with enhancement of the equatorward meridional winds and zonal wind shear. However, in the middle of the season, the day-to-day variations of PMSE OR have no relation to meridional winds nor to temperature. Only geomagnetic activity, expressed in local K-index, was found to correlate with them. We also found that inter-annual variations of PMSE OR and length of the season anticorrelate with solar activity, represented by the solar 10.7 cm radio flux, and correlate with geomagnetic activity, represented by Ap index. No statistically significant trends in PMSE occurrence rate or in the length of the PMSE season were found during 1997-2008.


In Paper II we concentrated on the accurate calculation of PMSE absolute strength as expressed by radar volume reflectivity. For this we introduced a new, independent calibration method, which can be used to estimate changes in transmitter output and antenna feed losses from year to year (for example due to changes in antenna configuration). This method is based on radar-radiosonde
comparisons in the upper troposphere/lower stratosphere region simultaneously with PMSE observations. Using this calibration, we calculated PMSE volume reflectivity from ESRAD measurements of PMSE during summers 1997-2009. We showed that the distribution of PMSE strength over magnitudes varies from year to year with the peak of the distribution varying from $2\times10^{-15}$ to $3\times10^{-14}$ m$^{-1}$. By using multi-regression analysis, we found that inter-annual variations of PMSE volume reflectivity strongly correlate with the local geomagnetic K-index and anticorrelate with solar 10.7 cm flux. The former correlation may be explained by an increase of electron density due to energetic particle precipitation, the latter effect may be explained by changes in the amount of water vapour and/or the mesospheric temperature. We did not find any statistically significant trend in PMSE volume reflectivity during 1997–2009.

**Paper III: Aspect sensitivity of polar mesosphere summer echoes based on ESRAD MST radar measurements in Kiruna, Sweden in 1997-2010.**

We used the long-term ESRAD PMSE data for calculation of in-beam PMSE aspect sensitivities using the FCA technique. We found that half of the PMSE detected each year are highly aspect-sensitive echoes and hence, cannot be explained by isotropic turbulence. The remaining half of the PMSE have an aspect-sensitivity parameter, characterising the half-width of the scatterers polar diagram, larger than 9-11°. We found that the distribution of PMSE over the aspect sensitivity parameter remains consistent from year to year with median values of 2.9-3.7°. The effect of such high aspect sensitivity can result in an underestimation of PMSE VR by more than a factor 2. PMSE aspect sensitivity shows altitude dependence with the scatter becoming more isotropic with increasing height. We did not identify any dependence of the PMSE aspect sensitivity on backscattered power for any year. After analysis of limitations of the in-beam and off-zenith beam methods we concluded that the former is suitable for highly aspect-sensitive echoes while the latter is needed for more isotropic scatterers.
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Bibliography


Polar mesosphere summer echoes with ESRAD, Kiruna, Sweden:
Variations and trends over 1997–2008

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1. Introduction

Polar Mesosphere Summer Echoes (PMSE) are strong radar echoes observed from mesopause altitudes at polar latitudes during summer time. A radio wave is scattered at irregularities in the radio refractive index when it reveals structures at the radar half wavelength (Tatarskii, 1961). At mesopause altitudes the radio refractive index is mainly determined by the free electron number density. It is now generally accepted that both charged ice particles and atmospheric turbulence play major roles in the creation of the electron number density structures that lead to PMSE in the mesopause region (Rapp and Lübken, 2004). From May to August in the northern hemisphere, temperatures of 130 K and lower are commonly found at altitudes 80–90 km. This allows particles to form from water vapor and to grow at mesopause altitudes (~85–88 km). Measured radii of these aerosols are typically between ~5 and 50 nm (Rapp and Lübken, 2004). The largest charged ice particles (larger than ~20 nm) are observed visually from the ground in the form of noctilucent clouds (NLC). Gravity waves propagating from the troposphere grow unstable in the mesosphere and create turbulence which produces irregularities in the radio refractive index, thus leading to radar echoes. More detailed reviews on PMSE have been published by Cho and Röttger (1997) and Rapp and Lübken (2004).

In recent years, NLCs have attracted considerable scientific interest because they have been suspected that these mesospheric clouds could be possible indicators of global change in the mesosphere (Thomas and Olivero, 2001). In practice, it is difficult to test this hypothesis because the long-term measurements of NLC are based mainly on visual and lidar observations which depend on human and weather factors. Recent statistics of the NLC occurrence rate (for NLC observations made by eye from the ground) show that there is no statistically confident overall long-term trend during the last 43 years (Kirkwood et al., 2007a). However, the authors commented that the observational limitations and high year-to-year variability in NLC makes it impossible to determine any trend less than 1% per year. PMSE observations with radar, unlike NLC observations, have the advantage of being continuous, and independent of observers and weather conditions. Thus PMSE observations over a long interval, at least over one solar cycle, are very attractive for studying variations and long-term trends in the atmospheric parameters at the polar mesopause.

PMSE were detected for the first time with the 50 MHz MST radar at the Poker Flat Research Range (65° N; 147° W), in Alaska, USA, in 1979 (Ecklund and Balsley, 1981). Since this first observation, PMSE measurements have been carried out by many radars, among others in the Northern Hemisphere—in Kiruna, Sweden (68° N; 21° E), Hankasalmi, Finland (62.3° N; 26.6° E), Anderes, Norway (69° N; 16° E), Svalbard, Norway (78° N; 16° E), Tromsø, Norway (69° N; 19° E), Resolute Bay, Canada (75° N, 95° W)—and in the Southern Hemisphere—in Antarctica at the Davis (69° S; 78° E), Wasa (73° S, 13° W) stations (e.g. Kirkwood et al., 1998; Ogawa et al., 2003; Bremer et al., 2009; Zecha and Röttger, 2009; Rapp et al., 2008; Huaman et al., 2001; Morris et al., 2007; Kirkwood et al., 2007b).

This paper presents the results of PMSE observations over the period 1997–2008 from the ESRAD 52 MHz radar located near Kiruna in Sweden. This is the world’s longest data set of PMSE...
observations made by the same radar at the same site with the same operating mode. We have analyzed the characteristics of the polar mesosphere summer echoes in terms of diurnal, day-to-day and year-by-year variations of occurrence rate and diurnal variations of the volume reflectivity. We have examined the possible relationship between day-to-day variations of PMSE occurrence rate, mesopause temperature, meridional wind and geomagnetic activity. Observations of winds and temperature, starting from 2003, were obtained with the collocated SKYMET meteor radar. We have investigated the changes in meridional wind and zonal wind shear associated with the start of the PMSE season. The influence of solar and geomagnetic activity on year-to-year variations in PMSE occurrence rate and the length of the PMSE season have been investigated, and the trends over whole 12-year interval have been studied.

2. Radars and data description

PMSE measurements have been carried out during the years 1997–2008 with the ESRAD MST (Mesosphere–Stratopause–Troposphere) radar in northern Scandinavia. (The year 1999 is not considered because of a radar malfunction.) ESRAD is a 52 MHz atmospheric radar located at the rocket range Esrange near Kiruna in Sweden (67.38° N; 21.10° E). The ESRAD antenna consists of an 18 x 16 (12 x 12 up to 2003) array of 5-element Yagis placed at 4.4 m (0.7 times the radar wavelength) from each other. This array is divided into 6 sub-arrays. Each sub-array is connected to a separate receiver. The main parameters of the ESRAD radar are shown in Table 1. A more detailed description of the ESRAD radar is given by Chilson et al. (1999). The radar is capable of operation in many different modes. The main radar modes usually used for long-term PMSE measurement are ‘fca_4500’ and ‘fca_150’. The first mode, ‘fca_4500’, uses an 8-bit complementary coded, 600 m resolution pulse train and a narrower receiver bandwidth, giving higher sensitivity. The second mode, ‘fca_150’, provides 150 m height resolution but has relatively poor signal-to-noise characteristics. The mode ‘fca_4500’ should be able to detect at least ten times weaker echoes than the mode ‘fca_150’. Therefore in our investigation we used only the mode ‘fca_4500’. The corresponding parameters are presented in Table 2. The height range of PMSE measurements was 80–90 km. Fig. 1 shows a typical example of PMSE observed by the ESRAD radar during one of the days in 2008.

Observations of winds and temperatures were obtained from the SKYMET (All-Sky Interferometric Meteor Radar) radar located less than 1 km from the ESRAD radar, at Esrange. SKYMET is a 32.5 MHz, multi-channel, coherent-receiver, pulsed radar. The receiving antenna consists of five separate crossed-element Yagi aerials, configured to act as an interferometer. A single aerial is used as the transmitter. The main parameters of the SKYMET radar are collected in Table 3. The radar has provided continuous data collection since August 5, 1999. For our investigation we selected data for hourly averaged winds from heights 80.8, 84.6, and 87.5 km because the mean PMSE altitude is about 85–86 km. The temperature measurements are evaluated from meteor decay time (Hocking, 1999). These temperatures are daily mean values at altitude range 87–90 km. The radar and the wind observations are described in more detail by Hocking et al. (2001) and Mitchell et al. (2002), respectively.

We calculated the volume reflectivity which provides us with information about the characteristics of the scatterers, independent of the characteristics of the radar used. Following Kirkwood et al. (2007b) we defined the volume reflectivity as

\[
\eta = \frac{P_t}{P_r} = \frac{1}{\pi D^2 F} 
\]

where \(P_t\) is power delivered to the radar, \(P_r\) is power received by the radar, \(D\) is the distance to the scattering volume, \(F\) is the range resolution along the radar beam, \(L\) is loss in the antenna feed (\(L=0.39\) was set), \(F\) is the fraction of the scattering volume which is filled with scatterers (\(F=1\) was assumed) and \(A_t\) is effective area of the receiving antenna \(A_t = \frac{L^2 C}{4\pi}\) where \(C\) is the antenna gain.

The power received by the radar can be estimated using the formula:

\[
P_r = k_B T_{sky} B_{filt} \left( S_{sys} + S_{sky} \right) C_{filt} C_{coh} C_{code} \]

where \(k_B\) is Boltzman’s constant, \(B\) is the receiver bandwidth, \(A_{filt}\) is the daily variation of the sky noise, \(S_{sys}\) is the total detected power, \(S_{sky}\) is the power due to internal system noise, \(S_{sky} + S_{sys}\) is the power due to sky noise, \(S_{sys}\) is the range of the daily variation in sky noise, \(C_{filt}\) is a receiver filter efficiency, \(C_{coh}\) the number of coherent integrations and \(C_{code}\) twice for the number of bits in the code. The range for \(A_{filt}\) is evaluated from assimilated maps of calibrated surveys of the radio-sky, scaled to 52 MHz (Oliveira Costa et al., 2008) and is given in Table 1. The sum of \(S_{sys} + S_{sky}\) was calculated by averaging the power detected at the range gates where scattered signal was not expected (30–50 km). Note that all of the parameters \(S, P\) were used in the same arbitrary units.

PMSE occurrence rates (OR) are calculated for every hour. OR is 1 if the volume reflectivity averaged over 1 h exceeds the

<table>
<thead>
<tr>
<th>Radar modes</th>
<th>fca_4500</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pulse length (3 dB)</td>
<td>2.1 μs</td>
</tr>
<tr>
<td>Sampling resolution</td>
<td>600 m</td>
</tr>
<tr>
<td>Filter bandwidth</td>
<td>128 (98.4 ms) per code up to 2004</td>
</tr>
<tr>
<td>Pulse repetition frequency</td>
<td>32 (24.6 ms) per code</td>
</tr>
<tr>
<td>Duty cycle</td>
<td>1300 Hz</td>
</tr>
<tr>
<td>Start altitude</td>
<td>2.7 μs</td>
</tr>
<tr>
<td>End altitude</td>
<td>4.8 km</td>
</tr>
<tr>
<td>Receiver filter efficiency</td>
<td>0.72</td>
</tr>
<tr>
<td>Mean attenuates for (S_{sys} + S_{sky})</td>
<td>30–50 km</td>
</tr>
</tbody>
</table>
threshold value of $10^{-15}$ m$^{-1}$ at any altitude in 80–90 km range. This threshold has been used because it is well above the background noise for ESRAD (Kirkwood et al., 2007b). More details on comparison of the results of using different volume reflectivity thresholds, in terms of day-to-day and year-by-year PMSE variations, are given in Sections 3.2 and 3.3.

In previous studies it was noticed that evaluation of $D_{S\text{sky}}$ on a daily basis can suffer from effects of interference and anomalous absorption of the sky noise (Kirkwood et al., 2007b). We find that the choice of the method to evaluation of $D_{S\text{sky}}$, either on a daily basis or using only selected quiet intervals without spikes, does not influence the diurnal variations of the PMSE volume reflectivity and OR averaged over the season.

3. Experimental results

3.1. Diurnal PMSE variations

Fig. 2 shows the results for diurnal variations of the logarithm of volume reflectivity averaged over the 80–90 km altitude range and the interval from 1 June to 5 August, for the different years. It is evident from the figure that the PMSE volume reflectivity has dominating semidiurnal variations with two maxima at 4–7 and 13–16 LT and two minima at about 7–11 and 20–24 LT. This tendency was found in all of the years studied, with the exception of 2004 and 2006 when only one broad maximum, at about 3–16 LT, and one evening minimum (20–24 LT) were observed. This evening minimum is the deepest part of the volume reflectivity curves for all years. We found that there are no significant changes in the shapes of the volume–reflectivity diurnal curves during the period from 1997 to 2008; however, the values of the maxima vary substantially between $10^{-14}$ and $10^{-12}$ m$^{-1}$ and the minima between $10^{-15.5}$ and $10^{-14}$ m$^{-1}$ range (note the logarithmic scale in the $y$-axis).

The results for the diurnal variations of PMSE occurrence rate averaged over the interval from 1 June to 5 August from the ESRAD observations for each year are shown in Fig. 3. The diurnal curves of PMSE occurrence rate show similar behaviour to those for volume reflectivity but their shapes look more irregular. The occurrence rates exhibit diurnal and semidiurnal variations and are characterized by maxima and minima observed almost at the same times as those for volume reflectivity. The signatures of semidiurnal variations are seen almost every year except 1997, 2003, 2004 and 2006. However because of irregular forms, diurnal variations can reveal several weak maxima and minima before the deep decrease which starts at about 16–18 LT. There are no significant changes of shape in the diurnal variations of occurrence rate curves during the period 1997–2008; however, their hourly mean values can change by factors 1.5–2.5 from year to year.

3.2. Day-to-day PMSE variations

The day-to-day variations of the PMSE occurrence rate observed by ESRAD for the all years studied are shown in Fig. 4 and details of seasonal characteristics are summarized in Table 4. The PMSE usually start in the last 10 days of May, increase sharply until the beginning of June, stay at the highest level until the end of July and then decrease gradually until the end of August every year. The earliest start of the PMSE season was detected on 19 May in 2003 and the latest one on 28 May in 2000. The average estimated time for the start of the PMSE season is 21–22 May. The earliest end of the PMSE season was 14 August 2000 and the latest one 26 August in 2004 and 2007. The average time for the end of PMSE season is 19–20 August. Finally the length of PMSE season lasts from the 79 days detected in 2000 up to 99 days in 2003.
A comparison of the day-to-day variations of PMSE occurrence rate using different volume–reflectivity thresholds is shown in Fig. 5. This figure illustrates day-to-day variations of PMSE occurrence rate from 1 June to 5 August for the year 2007, subject to three thresholds for the volume reflectivity: $2 \times 10^{-16}$, $10^{-15}$, and $5 \times 10^{-15}$ m$^{-1}$. The PMSE occurrence-rate curves are running means over 6 days to eliminate the possible influence of 5-day planetary waves (Kirkwood et al., 1998). Note that no significant changes in the shapes of the day-to-day curves for PMSE occurrence rate are found using the different thresholds.

3.2.1. Start of PMSE season in relation to meridional/zonal winds

Theory explains the extremely low temperatures at the summer mesopause, which are necessary for PMSE generation, as caused by dynamics—as a consequence of the gravity wave driven meridional circulation from the summer to winter mesosphere (e.g. Andrews et al., 1987). In order to test this experimentally we have investigated the change of meridional and zonal wind components associated with the start of the PMSE season. Daily mean meridional wind speed and zonal wind shear for different years are shown in Fig. 6. The measurements of wind were made by the SKYMET radar. The meridional wind was averaged at heights between 80.8 and
87.5 km. The zonal wind shear was calculated for the altitude range of 84.6–87.5 km. The time interval presented on the zonal wind shear was calculated for the altitude range of years 2000–2008. The mean components show the variations of meridional wind and zonal wind shear components averaged over years 2000–2008. The results show that, in the selected time interval, the mean meridional wind became more negative while the zonal wind shear became more positive. Positive and negative values of meridional wind correspond to the poleward and the equatorward component, respectively. Thus a few days before the start of the PMSE season, the mean meridional wind became more equatorward and the zonal wind shear grew.

This behaviour of the meridional wind is possible to see at the selected time intervals for all years. However, the zonal wind shear sometimes starts to increase more than 10 days after the start of the PMSE season (e.g. in years 2005 and 2008). In Fig. 7 we present the variations of meridional wind and zonal wind shear components averaged over years 2000–2008. The mean components show the same behaviour as was described for individual years but more clearly, despite their large variability from year to year.

### Table 4
The characteristics of PMSE seasons over 1997–2008.

<table>
<thead>
<tr>
<th>Year</th>
<th>The start of PMSE season</th>
<th>The end of PMSE season</th>
<th>The length of PMSE season</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997</td>
<td>23 May</td>
<td>18 August</td>
<td>88 days</td>
</tr>
<tr>
<td>1998</td>
<td>24 May</td>
<td>17 August</td>
<td>86 days</td>
</tr>
<tr>
<td>2000</td>
<td>28 May</td>
<td>14 August</td>
<td>79 days</td>
</tr>
<tr>
<td>2001</td>
<td>26 May</td>
<td>23 August</td>
<td>90 days</td>
</tr>
<tr>
<td>2002</td>
<td>27 May</td>
<td>22 August</td>
<td>88 days</td>
</tr>
<tr>
<td>2003</td>
<td>19 May</td>
<td>25 August</td>
<td>99 days</td>
</tr>
<tr>
<td>2004</td>
<td>23 May</td>
<td>26 August</td>
<td>96 days</td>
</tr>
<tr>
<td>2005</td>
<td>21 May</td>
<td>20 August</td>
<td>92 days</td>
</tr>
<tr>
<td>2006</td>
<td>20 May</td>
<td>23 August</td>
<td>96 days</td>
</tr>
<tr>
<td>2007</td>
<td>21 May</td>
<td>26 August</td>
<td>98 days</td>
</tr>
<tr>
<td>2008</td>
<td>25 May</td>
<td>22 August</td>
<td>90 days</td>
</tr>
</tbody>
</table>

3.2.2. Day-to-day PMSE OR variations in relation to geomagnetic activity, meridional wind and mesopause temperature

To investigate a possible connection between PMSE and geomagnetic activity we present the day-to-day variations of PMSE occurrence rate together with the local K index. Local K indices are based on the 3-h range of geomagnetic field values, as measured with flux-gate magnetometers at the Swedish Institute of Space Physics in Kiruna. We have also investigated the possible relation of PMSE occurrence rate to meridional wind and mesospheric temperature. We present day-to-day variations of PMSE OR, geomagnetic K indices, mean meridional wind and temperature for 6 years from 2003 to 2008 in Fig. 8. From the figure it can be seen that, in most cases, the PMSE occurrence rate curves have minima at about the same times as minima in geomagnetic activity have been detected, whereas the curves of mesospheric temperatures and meridional winds have no common features with PMSE occurrence rate curves.

In order to make a quantitative analysis of a possible correlation between the above-mentioned parameters, we have calculated the multi-parameter linear fit for the time interval from 15 June to 15 July for each year. This time interval was chosen to eliminate the influence of the start and the end of PMSE season. The fit is to equation:

\[
\text{OR} = A_0 + A_1 \times (T - T_{\text{mean}}) + A_2 \times (V - V_{\text{mean}}) + A_3 \times (K - K_{\text{mean}}).
\]

where OR is the daily mean occurrence rate of PMSE and \( T \), \( V \) and \( K \) are the daily mean values of the temperature, the meridional wind and the K index of geomagnetic activity respectively. \( T_{\text{mean}}, V_{\text{mean}} \) and \( K_{\text{mean}} \) are those mean values over the whole investigated time interval. For fitting we used the MATLAB “regress” function which calculates regression coefficient estimates and their 95% confidence intervals. Results of the fits are collected in Table 5. Due to a lack of temperature data, the result of the fit for the year 2006 is not presented.
The results show that in most cases there is an anti-correlation between PMSE occurrence rate and temperature but these values are not significant. The correlation between the occurrence rate of PMSE and mean meridional wind is not significant either and changes from positive to negative in different years. The positive correlation between occurrence rate of PMSE and $K$ index was found for all of the years included, but it is significant only for the years 2003 and 2007.

### 3.3. Long-term trend and variations

#### 3.3.1. Year-to-year variations of PMSE occurrence rate

The year-to-year variations of PMSE occurrence rate for different volume reflectivity thresholds are shown in Fig. 9. This figure illustrates the PMSE occurrence rate averaged over the time interval from 1 June to 5 August for the years 1997–2008 subject
to three thresholds for the volume reflectivity: $2 \times 10^{-16}$, $10^{-15}$, $5 \times 10^{-15}$ m$^{-1}$. Note that we do not find significant changes in the shapes of the year-to-year curves for PMSE occurrence rate using the different thresholds.

The upper panel of Fig. 10 shows the year-to-year variations of the PMSE occurrence rate together with mean values of geomagnetic and solar activity for period from 1 June to 31 July during the years 1997–2008. The solar activity is represented by the
solar 10.7 cm radio flux and the geomagnetic activity is presented by the $Ap$ index. $Ap$ is a planetary index based on the $K$ indices obtained at 13 mid-latitude geomagnetic observatories located all over the globe. We chose it here instead of the local $K$ index because we want to compare our results with a previous long-term study by Bremer et al. (2009) where the $Ap$ index was used.

From investigation of long-term changes in noctilucent clouds it is known that it is necessary to allow for a quasi-decadal variation, which may be due to the solar activity cycle (Kirkwood et al., 2007a). In order to eliminate the possible influence of solar flux and geomagnetic activity to the PMSE occurrence rate we have applied multi-parameter regression analysis. We used the multi-parameter linear fit with all possible predictors such as constant, trend, solar flux and $Ap$ index of geomagnetic activity. Thus the fit is to equation:

$$\text{OR} = A_0 + A_1 \times (year - 1997) + A_2 \times (F_{mean} - F_{mean}) + A_3 \times (Ap_{mean} - Ap_{mean})$$

(4)

where OR is the yearly occurrence rate of PMSE averaged over the period from 1 June to 31 July for each year. $F$ and $Ap$ are the mean solar 10.7 cm flux and the $Ap$ index of geomagnetic activity, respectively, over the same interval. $F_{mean}$ and $Ap_{mean}$ are those mean values over the years 1997–2008. For fitting we used the MATLAB “stepwise” function which calculates the regression coefficient estimates, their 95% confidence intervals, probability for testing whether coefficients are 0 ($p$-values). The result of the multi-parameter linear fit is presented in Table 6.

The obtained regression coefficients show that there is no significant dependence between the variations in PMSE occurrence rate from year to year and solar activity. However a significant, positive correlation is found between PMSE occurrence rate and geomagnetic activity. After elimination of the influence of solar and geomagnetic activity a possible long-term trend can be found using:

$$\text{OR} = \text{OR} - A_1 \times (F_{mean} - F_{mean}) - A_2 \times (Ap_{mean} - Ap_{mean}) = A_0 + A_1 \times (year - 1997)$$

(5)

The lower panel of Fig. 10 presents the year-to-year variations of PMSE occurrence rate (after correction for solar and geomagnetic activity terms) together with the fitted trend. The year-to-year variations of PMSE occurrence rate show a negative trend of about 2.4% for the whole period 1997–2008 but this value is not significant (66% probability that the trend is zero).

![Figure 9](image9.png)

**Fig. 9.** Year-to-year variations of PMSE occurrence rate for different thresholds at volume reflectivity: $2 \times 10^{-18}$ m$^{-1}$ (dashed line), $10^{-18}$ m$^{-1}$ (solid line), $5 \times 10^{-18}$ m$^{-1}$ (dash-dotted line).

![Figure 10](image10.png)

**Fig. 10.** Upper panel: year-to-year variations of PMSE occurrence rate (solid line), $Ap$ index (dash-dotted line) and solar flux (dashed line). Lower panel: year-to-year variation of the PMSE occurrence rate (solid line), after elimination of solar and geomagnetic activity influences, and its trend (dashed line).
3.3.2. Year-to-year variations of the length of PMSE season

Fig. 11 (upper panel) shows the year-to-year variations of the length of the PMSE season for the years 1997–2008 together with mean values of geomagnetic and solar activity for period from 1 June to 31 July. We have applied the same multi-parameter regression analysis as for OR to the variations of PMSE season length. We have calculated a linear fit to the equation:

\[ L = A_0 \times \text{year}^{-1} + A_2 \times F_{\text{mean}} + A_3 \times Ap_{\text{mean}}. \]

where \( L \) is the length of PMSE season, and other notations are the same as for the linear fit for OR.

The result of the multi-parameter linear fit is presented in Table 6. It shows that there is a significant negative correlation between the year-to-year variation in the length of PMSE season and solar activity. We have also found a positive correlation between the length of the PMSE season and geomagnetic activity. In the same way as before we have eliminated the terms of solar and geomagnetic activity from the year-to-year variations of PMSE season length and derived its trend (see Fig. 11, lower panel). Finally we found a positive trend about 5 days in variations of length of the PMSE season for the whole period from 1997 to 2008; however this value is not significant.

### Table 6

Coefficients of multi-parameter regression for year-to-year variation of OR and length of the PMSE season, \( L \), together with their 95% confidence intervals. P-values give the probability that the coefficient is zero. The last column (\( p \)) is the probability that all coefficients are zero for the final fit with minimum root mean square error when coefficients without parentheses are included.

<table>
<thead>
<tr>
<th></th>
<th>( A_0 )</th>
<th>( A_0/\text{year} )</th>
<th>( A_2/F )</th>
<th>( A_3/\text{Ap} )</th>
<th>( p )</th>
</tr>
</thead>
<tbody>
<tr>
<td>OR</td>
<td>0.72 ± 0.09</td>
<td>(−0.002 ± 0.015)</td>
<td>−0.0009 ± 0.013</td>
<td>0.0107 ± 0.0102</td>
<td>0.11</td>
</tr>
<tr>
<td>L days</td>
<td>89 ± 5</td>
<td>0.44 ± 0.81</td>
<td>−0.103 ± 0.083</td>
<td>0.681 ± 0.018</td>
<td>0.02</td>
</tr>
</tbody>
</table>

![Fig. 11. Upper panel: year-to-year variations of the length of PMSE season (solid line), Ap index (dash-dotted line) and solar flux (dashed line). Lower panel: year-to-year variation of the length of PMSE season (solid line), after elimination of solar and geomagnetic activity influences, and its trend (dashed line).](image)

4. Discussion

4.1. Diurnal PMSE variations

Making use of a data set covering 11 PMSE seasons, we have found that both PMSE strength (volume reflectivity) and occurrence rate show semi-durnal variations with two maxima at 4–7 and 13–16 LT, a deep minimum at 20–24 LT and secondary minimum at 7–11 LT. These results agree well with previous studies. Barabash et al. (1998) obtained the same results for just 1 year (1997) of PMSE observations with ESRAD. A similar pattern of PMSE SNR (signal-to noise ratio) daily variations was found with the ALOMAR VHF radar for 1994–1997 (Hoffmann et al., 1999) and for PMSE OR for 2004 (Latteck et al., 2007). On the basis of one season 2004–2005 of observations Morris et al. (2007) reported about similar semi-diurnal variations of PMSE strength and OR at Davis, Antarctica. Some discrepancies in determination of the temporal positions of maxima and minima can be due to irregularity of the shape of the diurnal curves, which change their positions from year to year, and finally due to different quantities such as reflectivity in the present case and SNR in other cases being used for analysis.

Whereas the existence of the diurnal and semi-diurnal variations has been reported by many scientists, the mechanism behind them is not completely clear yet. PMSE intensity can be affected by variations of electron density and temperature at the same heights.
(see, e.g., formula in Rapp et al., 2008 where temperature affects electron diffusivity and hence the Schmidt number). However, it is clear that their contributions are different. Kloeppermeier (1999) fitted a simple model for diurnal variation of the radar reflectivity to the observations which includes tidal temperature variation, electron production by solar ionization and by energetic particle precipitation. He concluded that the two latter factors mainly determine the diurnal variation of reflectivity and the contribution of the temperature variation is small. Hoffmann et al. (1999) found a significant but small anti-correlation of about 0.25 between PMSE SNR and 3-h-shifted meridional wind tides which should be associated with tidal temperature variations. This implies that only 6.25% of total SNR variability can be accounted by that in temperature. Probably the rest of SNR variation is determined by changing ionization and other factors.

At polar latitudes the ionization of the lower ionosphere is mainly due to the Lyman α band of solar radiation and medium-energy particle precipitation from the magnetosphere. The former has a pronounced maximum at local noon while the latter has a deep minimum at 14–19 MLT depending on geomagnetic activity (Codrescu et al., 1997). Thus a superposition of these two ionization sources might explain diurnal variations of PMSE strength (Morris et al., 2005; Nilsson et al., 2008).

In turn, the PMSE OR depends on electron density indirectly, via the given threshold. A threshold defined as a minimum detectable reflectivity is affected mainly by variations in ionization and to smaller extent by those in temperature as discussed above. From Fig. 5 we see that the choice of the threshold does not influence the shape of the seasonal curves for OR. Thus one can also expect that the OR during the day will be modulated by diurnal variations of the threshold caused mainly by variations in electron density. As a result the diurnal variations of PMSE reflectivity and OR are rather similar as demonstrated in Figs. 2 and 3.

We have found the shapes of the PMSE reflectivity daily curves do not change much from year to year, while amplitudes of maximum and minimum vary up to 1.5–2 orders of magnitude during the 12-year interval. This might imply that the mechanism responsible for such variations remains the same, whereas overall level of ionization changes on a scale of one solar cycle. That is not surprising because electron density at 80–90 km altitudes under geomagnetically quiet conditions (i.e. defined by solar radiation only) has been found to vary up to an order of magnitude between solar minimum and maximum (Kirkwood, 1993). Aminaei et al. (2006) have showed that occurrence rate and amplitude of spikes in riometer data, which are associated with auroral particle precipitation, change by 14% and 1 dB, respectively, from solar minimum to solar maximum. Because the diurnal curves in Fig. 2 do not show systematic behaviour over the course of one solar cycle, we suggest that other factors besides ionization rate can play role. For example, the concentration of water vapor and availability of ice nuclei can vary from year to year. Similar analysis for long-term variability of the diurnal variation of PMSE OR is complicated because OR is less sensitive to electron density than the volume reflectivity.

4.2. Day-to-day PMSE variations

4.2.1. Seasonal behaviour

From the first observations of PMSE carried out at Poker Flat, Alaska (65 N), between February 1979 and December 1980, it has been known that PMSE shows pronounced seasonal variations (Edkins and Balley, 1981). We have presented a more detailed climatology related to the day of start, end and duration of PMSE season (see Table 4). These results confirm previous shorter-term observations of PMSE with ESRAD in 1997 (Kirkwood et al., 1998) and agree with recent long-term observations of PMSE at Andenes during the years 1999–2008 reported by Bremer et al. (2009). Note that the determination of days and time range depends on the selection of the threshold for volume reflectivity and radar characteristics and therefore might vary slightly from one radar to another.

4.2.2. Start of PMSE season in relation to meridional/zonal winds

It is generally believed that the very cold temperatures at the summer mesopause result from forcing of the meridional circulation by gravity waves. Gravity waves propagating from the troposphere include meridional and zonal propagating modes. Zonal propagating modes can be eastward or westward relative to the mean flow. In the summer time, when westward winds exist in the stratosphere, gravity waves with eastward phase speed can reach the mesosphere and break there. These breaking gravity waves cause a deceleration of the zonal wind and, by thermal wind balance, a meridional temperature gradient. A residual circulation, with upwelling of the air mass at the summer pole and downwelling at the winter pole, with zonal-mean meridional flow from the summer to the winter hemisphere at mesopause heights, maintains this temperature distribution. Thus in the northern hemisphere during summer an equatorward wind is associated with the transport of cold air and a poleward wind with the transport of warm air.

We found that a few days before the start of the PMSE season the meridional wind becomes more negative, which corresponds to the equatorward component. It also intensifies. At the same time, the amplitude of zonal wind shear increases. This is clearly seen for mean values (Fig. 7) and also for majority of individual years (Fig. 6). Thus, our results agree with generally accepted theory and allow us to make the conclusion that the start of the PMSE season is associated with an enhancement of the equatorward winds.

4.2.3. Day-to-day PMSE variations in relation to mesopause temperature and meridional winds

The anomalous cold temperatures in the mesosphere are responsible for the formation of ice particles from water vapor. Thus investigations of a possible connection between the day-to-day variations of PMSE occurrence rate and mesospheric temperature are important for PMSE study. Day-to-day PMSE OR can be strongly influenced by planetary waves (e.g. Kirkwood et al., 2002; Klekociuk et al., 2008). Variations in PMSE occurrence with a 4–6 day period were observed by the ESRAD radar during the summer of 1997 (Kirkwood and Rechou, 1998). The authors found that at the beginning and end of the PMSE season, they are closely anti-correlated with temperature variations associated with 5-day planetary waves extracted from the U.K.M.O assimilated global data analyses. The first simultaneous observations of temperature with the SKYMET radar and of PMSE with ESRAD in the summer of 2000 demonstrated the sensitivity of PMSE to temperature fluctuations associated with 5-day waves and confirmed that temperature is the primary parameter affecting PMSE occurrence (Kirkwood et al., 2002). However in this paper we aim to find which factors other than planetary waves can affect PMSE behaviour from day to day.

Recent investigations using satellite data for mesospheric temperature reported a causal linkage between meridional wind and the thermal structure of the mesopause region which is reflected in modulation of the PMSE OR envelope at Davis, Antarctica (Morris et al., 2007, 2009). However, this modulation may rather be associated with 5-day planetary waves than with day-to-day variations. Zeller et al. (2009) have studied a possible correlation between temperature and PMSE SNR from the ALWIN radar at Andenes for two seasons, 2002 and 2003. The authors
reported significant negative correlation during the PMSE season in 2002, when the mesospheric temperature was anomalously high and close to the water vapor frost point. The dependence between temperature and PMSE variations in 2003 was described as follows: at the beginning and end of the season the correlation is significant and strongly negative but in the middle of the season the correlation is weaker and not significant. The authors concluded that temperature variations have only a low impact on PMSE variations when the temperature is significantly below the frost point.

We have investigated relationships between PMSE occurrence rate and temperature during six PMSE seasons from 2003 to 2008 using regression analysis. We have found that day-to-day variations for the interval from 15 June to 15 July are negatively correlated with the temperatures in all investigated years with one exception, 2008. However, during this interval when PMSE stayed at the highest level the anti-correlation is not significant. Hence the low temperatures which lead to PMSE formation do not influence the variations during the middle of the season. Thus with this 6-year data set we have confirmed the result obtained for 1 year (2003) by Zeller et al. (2009). Note that this result must be considered with some caution because of some difference in the heights of observation of the temperatures (−90 km) and PMSE layers (−86 km) as was described in Section 2.

Using regression analysis for ESRAD and SKYMET radar data we did not find any significant correlation between day-to-day variations of PMSE occurrence rate and meridional wind in the time interval 15 June–15 July. Thus it seems that the meridional wind affects the formation of PMSE but does not influence their day-to-day variation in the middle of the season.

Thus the dependence between temperatures and PMSE variations can be described as follows: low temperatures (lower than water vapor frost point) would be needed in May to allow ice particles to form from water vapor and produce PMSE but later, in the middle of the season, if planetary waves with large amplitudes are absent, the temperatures remain below the frost point and do not influence the day-to-day variations of PMSE directly. The equatorward meridional wind associated with the temperature decrease leading to the start of the PMSE season, does not influence day-to-day PMSE variations in the middle of season.

4.2.4. Day-to-day PMSE variations in relation to geomagnetic activity and cosmic noise absorption

We found a significant positive significant correlation between the day-to-day variations of PMSE occurrence rate and K index for the years 2003 and 2007 in the time interval from 15 June to 15 July (see Table 5). In order to check the relationship between ionization in response to energetic particle precipitation and PMSE OR, we have tested the correlation between the day-to-day PMSE variations and cosmic noise absorption (CNA) data from a riometer in Abisko, which is about 100 km north-west from the ESRAD site (http://www.sgo.fi/Data/Riometer). CNA data were available for the years 2000–2004 and we used the same time interval from 15 June to 15 July as for the K index. We found small correlations of both signs between PMSE OR and CNA (not shown here).

Here we should discuss differences between magnetic indices, cosmic noise absorption (CNA) and electron density in the lower ionosphere, differences that are often forgotten or simplified. K and Ap geomagnetic indices are based on geomagnetic field measurements and as such depend on both ionospheric conductivities (and hence electron density) and electric field mainly above 100 km altitude. In turn, CNA is determined by (height-integrated) electron density mainly below 100 km. Recently Kellerman et al. (2009) reported that the horizontal component of magnetic perturbations in the auroral region is highly correlated with CNA only during the 3 h after magnetic local midnight (which is about 1 h before solar midnight in Kiruna). At other times the correlation is poor, resulting in a daily averaged value of 0.4 (our estimation). This is close to the correlation coefficient of 0.5–0.7 which we find between daily mean values of the Kiruna K index and CNA measured by the riometer in Abisko for the years 2000–2004 (not presented here). Kellerman et al. (2009) showed also that correlation between magnetic perturbations and ionospheric electric field is higher than that between the former and CNA. This might imply that the K and Ap indices are more likely associated with ionospheric electric field than with electron density variations below 100 km. And the latter is represented by CNA variations.

There have been several studies dedicated to the effect of magnetic activity and CNA on PMSE strength and only a few on PMSE OR. Morris et al. (2005) reported weak correlation (0.3–0.4) between 1-h averaged data for CNA and PMSE SNR measured at the Antarctic site Davis during the summer of 2003–2004. In contrast, Barabash et al. (2002) did not find significant correlation between day-to-day variations in PMSE intensity and CNA, comparing the same time on each day using 1-min resolution data (for the summer of 1997). Barabash et al. (2004) made a case study of the strong solar proton event (SPE) on July 15, 2000 and its whole effect on the lower thermosphere, mesosphere and PMSE. They found evidence that enhancement of the ionospheric electric field rather than electron density and neutral temperature might explain the observed disappearance of PMSE during part of the SPE. Strong electric fields associated with energetic particle precipitation can produce vertical transport of charged aerosols away from the PMSE region which might violate the conditions for PMSE generation.

Most recently Zeller and Bremer (2009) have presented results of a superimposed epoch analysis which shows that PMSE OR measured by the ALWIN MST radar in Andenes for the years 1999–2005 were stronger on the 1–2 days after the maximum Ap index for moderate geomagnetic activity. However the significance level of the result was relatively low. With higher significance Zeller and Bremer (2009) showed that PMSE OR decreases on the day of local K index minimum. Thus we can summarize that for low and moderate geomagnetic activity a positive correlation was found between geomagnetic indices and PMSE OR. Zeller and Bremer (2009) proposed the reason for the PMSE enhancement to be increasing electron density caused by high energetic particles precipitating down into the lower ionosphere during and after the geomagnetic disturbance. They also found that for 9 strong magnetic disturbances the effect is the opposite: PMSE OR decreases on the same day and one day after. This result is in perfect agreement with the complete disappearance of PMSE on July 15, 2000, during strong SPE when the daily mean Ap was 400 (Barabash et al., 2004).

We cannot directly compare the results by Zeller and Bremer (2009) obtained on an event basis to our results. We used 30-day time series for K index, CNA and PMSE OR, where days with low K index as well as with high K index were present. Thus, our result reflects some averaged situation. Therefore we could just speculate that when K index is large, the electric field is large too and the mechanism proposed by Barabash et al. (2004) might work to explain the decrease of OR found by Zeller and Bremer (2009). If K index is small, so is the electric field, and the dependence of OR on electron density below 100 km (i.e. on CNA), as discussed in Section 4.1, starts playing a role.

4.3. Long-term trend and variations

4.3.1. Year-to-year PMSE variations in relation to solar and geomagnetic activity

We have found negative correlation between the year-to-year variations of PMSE (both occurrence rate and length of the season)
and solar activity, represented by the solar 10.7 cm radio flux. In turn there is a positive correlation between long-term PMSE variations and the Ap index of geomagnetic activity. In contrast there is a positive correlation between long-term PMSE and solar activity, represented by the solar 10.7 cm radio flux. In season duration; however the significance levels of both trends 2008 revealed positive trends in PMSE occurrence rate and PMSE Previous investigations of long-term trends in PMSE carried out at 4.3.2. Trends in PMSE occurrence rate and length of PMSE season can be applied here. The relationship of PMSE to solar activity might not be so simple and direct. Increased solar activity leads to extra ionization in the lower ionosphere, i.e. at PMSE altitudes, and to growth of PMSE OR as was discussed in Section 4.1 of this paper. However increasing solar radiation might also increase the mesospheric temperature and increase photo-dissociation of water vapor which should give a negative correlation with PMSE OR. Thus the net solar effect is difficult to predict. Interestingly, NLC occurrence recorded by European observers for 43 summer seasons since 1964 shows a good anti-correlation with the 10.7 cm solar flux but with a lag of 13–17 months (Kirkwood et al., 2007a). Therefore it cannot be explained by a direct solar influence on NLC; changes in stratospheric circulation due to cycling of solar radiation were suggested by Kirkwood et al. (2007a) as a possible explanation. We have made a cross-correlation analysis between yearly PMSE OR and the 10.7 cm solar flux, but the best (but still low) anti-correlation was obtained for zero lag unlike for NLC. So far we have no explanation for this phenomenon. Regarding the relation between yearly PMSE OR and geomagnetic activity, this result is similar to that obtained for day-to-day variations of these quantities. Thus, the discussion in Section 4.2.4 can be applied here. 4.3.2. Trends in PMSE occurrence rate and length of PMSE season We did not find any significant trends in PMSE occurrence rate and the length of the PMSE season during the years 1997–2008. Previous investigations of long-term trends in PMSE carried out at Andenes with two different VHF radars at 53.5º Hz during 1994–2008 revealed positive trends in PMSE occurrence rate and PMSE season duration; however the significance levels of both trends are low (Bremer et al., 2009). Our results agree also with results obtained by Kirkwood et al. (2007a) for trends in NLC occurrence observed from the UK and Denmark over 43 years. The authors reported that no evidence is found for a statistically significant trend in the length of the NLC season, nor in the number of NLC nights when moderate and bright clouds were observed. Analysis of another NLC data set from Moscow for 1962–2005 revealed some positive but non-significant trend for NLC occurrence (Dalin et al., 2006). The absence of significant trends in PMSE and NLCs is in some contrast to results based on satellite measurements of polar mesospheric clouds (PMCs). Using 28 years of satellite observations of PMCs, Shettle et al. (2009) reported an increase in their occurrence frequency by 7% per decade at 64–74º latitudes. We can suggest at least two possible reasons for disagreement in long-term trends in PMSE and PMCs. First of all our observations of PMSE are for a shorter time, and therefore the trends obtained are below the significance level. As a result the differences in OR for only 2 years 2004 and 2006 are able to change the trend sign from positive (Fig. 8 of Bremer et al., 2009) to negative (Fig. 9, this article). It means also that there is a small but finite chance that the PMSE OR trend is as high as that found for PMCs. Another reason is that PMSE and PMCs are most sensitive to different parameters. PMSE OR may be mainly affected by electron density as was discussed in Section 4.1. On the other hand PMC trends may reflect those in the mesopause temperature and water vapor content. 5. Conclusion We have used a unique long data set of PMSE seasons (1997–2008) of calibrated observations by the ESRAD radar at Kiruna to come to the following conclusions. The PMSE occurrence rate and volume reflectivity on a daily scale show semidiurnal variations with two maxima (4–7 and 13–16 LT) and two minima (7–11 and 20–24 LT). Not much interannual variation is seen in the shape of the diurnal curves for occurrence rate and reflectivity; however, the absolute value of the minimum in occurrence rate varies from 0.2 to 0.7, and the maximum from 0.7 to 0.9, between different years. The PMSE season usually starts around May 21–22, reaches its peak by the beginning of June, remains at the highest level until the end of July and then gradually decreases towards the end of August every year. The length of the PMSE season varies from year to year (79 days in 2000 to 99 days in 2003). The start of the PMSE season is associated with enhancement of the equatorward meridional winds and zonal wind shear. However, neither the meridional wind nor the temperature, which are anti-correlated with the start of the PMSE season, seem to influence day-to-day PMSE variations in the middle of the season. At the same time day-to-day variations of geomagnetic activity are correlated with those of PMSE occurrence rate, when the occurrence rate is at its highest level in the middle of the season. We found indications that both ionospheric electric field and electron density may affect PMSE OR fluctuations on a day-to-day basis. The correlation of the year-by-year variations of PMSE occurrence rate and length of season with solar activity, represented by the solar 10.7 cm radio flux, is negative but not significant. Geomagnetic activity is one of the factors which might be responsible for interannual variations of PMSE occurrence rate and length of PMSE season. We did not find any significant trends in PMSE occurrence rate and length of PMSE season during the years 1997–2008. The choice of threshold for PMSE detection does not change the shape of either the seasonal or the long-term curves of PMSE occurrence rate. Therefore the results obtained in this paper, e.g. for the absence of a long-term trend in PMSE, do not limit themselves to the given threshold but can be considered as general. Acknowledgements ESRAD is a joint venture between Swedish Institute of Space Physics and Swedish Space Corporation, Esrange. Work by Maria Smirnova was funded by the Swedish National Graduate School of Space Technology hosted by Luleå University of Technology. References Aminaei, A., Honary, F., Kavanagh, A.J., Spannwick, E., Viljanen, A., 2006. Characteristics of night-time absorption spire events. Ann. Geophys. 24, 1887–1904. Andrews, D.G., Holton, J.R., Leovy, C.B., 1987. Middle Atmospheric Dynamics. Academic Press, New York and London. Barabash, V., Chilson, P., Kirkwood, S., Rechou, K., Stebel, K., 1998. Investigations of the possible relationship between PMSE and tides using a VHF MST radar. Geophys. Res. Lett. 25 (17), 3297–3300. Barabash, V., Chilson, P., Kirkwood, S., 2002. Are variations in PMSE intensity affected by energetic particle precipitation? Ann. Geophys. 20, 539–545. Barabash, V., Fedorov, A., Kirkwood, S., Kuderov, A., 2004. Polar mesosphere summer echoes during the July 2000 solar proton event. Ann. Geophys. 22, 759–771. Bremer, J., Huffman, P., Latchock, R., Singer, W., Zecha, M., 2000. Long-term changes of (polar) mesosphere summer echoes. J. Atmos. Solar Terr. Phys. 71, 1571–1576.
Polar mesosphere summer echo strength in relation to solar variability and geomagnetic activity during 1997–2009

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Abstract. This paper is based on measurements of Polar Mesosphere Summer Echoes (PMSE) with the 52 MHz radar ESRAD, located near Kiruna, in Northern Sweden, during the summers of 1997–2009. Here, a new independent calibration method allowing estimation of possible changes in antenna feed losses and transmitter output is described and implemented for accurate calculation of year-to-year variations of PMSE strength (expressed in absolute units – radar volume reflectivity \( \eta \)). The method is based on radar-radiosonde comparisons in the upper troposphere/lower stratosphere region simultaneously with PMSE observations. Inter-annual variations of PMSE volume reflectivity are found to be strongly positively correlated with the local geomagnetic K-index, both when averaged over all times of the day, and when considering 3-h UT intervals separately. Increased electron density due to energetic particle precipitation from the magnetosphere is suggested as one of the possible reasons for such a correlation. Enhanced ionospheric electric field may be another reason but this requires further study. Multi-regression analysis of inter-annual variations of PMSE \( \eta \) shows also an anti-correlation with solar 10.7 cm flux and the absence of any statistically significant trend in PMSE strength over the interval considered (13-years). Variations related to solar flux and K-index account for 86% of the year-to-year variations in radar volume reflectivity.

Keywords. Meteorology and atmospheric dynamics (Middle atmosphere dynamics) – Radio science (Instruments and techniques)

1 Introduction

Since the 1980s it has been known that extremely strong radar echoes can be detected from altitudes of 80–90 km, in high latitude regions, during summer. These echoes are known as Polar Mesosphere Summer Echoes (PMSE) and they are closely related to the visible phenomenon known as noctilucent clouds (NLC). A recent review of PMSE can be found in Rapp and Lübken (2004). While visible NLC indicate the presence of ice particles with sizes of several tens of \( \mu \)m (Hervig et al., 2001; Gumbel et al., 2001) formed at extremely low summer mesopause temperatures (\( \sim 130 \) K), PMSE occur due to scattering from fluctuations in electron density caused by a combination of electrically charged ice particles, including those of smaller sizes, and atmospheric turbulence. Thus PMSE have a dual nature: they are related both to ionospheric plasma and to the neutral atmosphere, in particular the thermal and dynamical state of the mesosphere.

Since NLC are sensitive to temperature and water vapour concentration (e.g. Lübken et al., 2007), it has been proposed that they might serve as an indicator of the values of these parameters in the mesosphere and be useful for detection of possible composition changes due to changing climate (Thomas and Olivero, 2001). Thus possible long-term trends in NLC derived using both ground-based and satellite-borne measurements attract great interest and discussion in the literature (e.g. DeLand et al., 2003, 2007; Kirkwood et al., 2008; Shettle et al., 2009).

PMSE are readily measured by VHF radars and have been monitored on a continuous basis since the 1990s. Therefore PMSE can also be used for long-term variability studies (e.g. Bremer et al., 2009; Smirnova et al., 2010). PMSE observations with radars have advantages over ground-based visual and lidar observations of NLC because they are weather and observer independent, and can be made 24 h a day. Satellite measurements of NLC are free from weather effects but have to be corrected for differences in observation times (DeLand
et al., 2007), and there is still some uncertainty in these corrections. However, when analysing secular variations and trends in PMSE one should bear in mind their dual nature, which means they are influenced not only by the characteristics of ice particles (NLC) but also by variations in the ionospheric plasma, which are caused by solar and geomagnetic variability. There is, for example, a clear positive correlation between the index of geomagnetic activity and PMSE occurrence rate (Bremer et al., 2006; Smirnova et al., 2010). In some studies a PMSE detection threshold value is set for signal-to-noise ratio (SNR), in others a threshold value for the radar volume reflectivity \( \eta \) is used. However, even the use of the same threshold in terms of SNR, for radars with different noise levels, implies different signals, i.e. PMSE strengths. Therefore, OR calculated according to this procedure is not suitable for comparative studies of PMSE. Unlike SNR, the radar volume reflectivity is an intrinsic characteristic of PMSE strength which can be used for comparison of PMSE measured with different radars at different locations. An additional advantage of using \( \eta \) is in the fact that its value should be directly related to physical parameters such as electron and ice particle density, static stability and turbulence characteristics (Kirkwood et al., 2010b; Varney et al., 2010). Nowadays estimations of PMSE reflectivity have been made using a number of Mesosphere-Stratosphere-Troposphere (MST) radars located both in the northern and southern polar regions (e.g. Kirkwood et al., 2007; Latteck et al., 2007, 2008; Swarnalingam et al., 2009). Comparisons between sites are dependent on the use of accurate calibration methods. So far, slightly different methods of radar calibrations have been applied at different sites which lead to uncertainty in the accuracy of cross-comparisons.

This paper is based on observations of PMSE during 1997–2009 using the MST radar ESRAD located near Kiruna in Northern Sweden. ESRAD has been running continuously since 1997 and has measured PMSE every summer using the same experimental parameters. Here we introduce a new method of independent calibration which allows accurate comparison of radar sensitivity from year to year for the ESRAD radar and can in principle be used for improving the accuracy of comparisons between sites. By using this unique PMSE data set, together with the new method of calibration, we can accurately calculate year-to-year variation of PMSE strengths (expressed in absolute units – radar volume reflectivity) over an interval exceeding one solar cycle. An important advantage of this study is that the measurements of PMSE were carried out with the same radar at the same place. This makes the data set from ESRAD more attractive for long-term study than other observations.

The goal of this paper is (1) to accurately calculate the inter-annual variations of radar volume reflectivity for PMSE over Kiruna, Sweden; (2) to quantify the contributions of solar and geomagnetic activity to those variations and (3) to test for any secular trend in PMSE strength. We start with a description of the radar and the new method for independent calibration. Then yearly averaged radar volume reflectivities for PMSE are calculated with correction according to the calibration. Finally, the relationships between year-to-year variations of PMSE \( \eta \), solar 10.7 cm flux and local geomagnetic K-index are evaluated.

2 PMSE strength over summers 1997–2009

2.1 Radar and experiment description

PMSE have been observed with the ESRAD MST radar, located near Kiruna, Sweden (67.88° N, 21.10° E), during each summer 1997–2009, although observations from 1999 are not included here because of a radar malfunction. ESRAD operates at a frequency of 52 MHz. The transmitter has 72 solid-state modules each of 1 kW, thus the resulting peak power reaches 72 kW with a maximum duty cycle of 5%. The pulse repetition frequency can vary from 100 Hz to 16 kHz and the pulse lengths from 1–50 \( \mu s \). The radar is able to use both Barker and complementary codes for pulse coding of the transmitted signals. The sampling interval of the receiving system can be between 1 \( \mu s \) and 20 \( \mu s \).

The radar provides information on the dynamic state of the atmosphere such as winds, waves, turbulence and layering, from the troposphere up to the lower stratosphere (ca. 1 km–20 km altitude) and at higher altitudes when strong scatterers are present (e.g. PMSE). A detailed descriptions for ESRAD radar is given by Chilson et al. (1999), updated by Kirkwood et al. (2007).

Initially the ESRAD antenna consisted of 12 × 12 phased array of 5-element Yagis spaced at 4.04 m (0.7 times the radar wavelength) from each other. In April 2004 an extended antenna array (constant phase) with 16 × 18 Yagis was taken into use. The array is divided into 6 sub-arrays, each connected to separate receivers, to detect backscattered power from the atmosphere. During the PMSE measurements used here, ESRAD ran an operating mode ‘\( f_{\mu a 4500} \)’ which provides an 8-bit complementary code, 600 m resolution, pulse train and a narrow receiver bandwidth, resulting in high sensitivity. The parameters of this experiment are presented in Table 1.

2.2 Calculation of radar volume reflectivity

Radar volume reflectivity \( \eta \) can be calculated according to the formula (e.g. Gage, 1990):

\[
\eta = \frac{P_r 64 (2\ln 2)^2}{P_i \pi L A_v \Delta r}
\]

(1)

where \( P_r \) is power delivered to the radar, \( P_i \) is power received by the radar, \( r \) is the distance to the scattering volume, \( \Delta r \) is the range resolution along the radar beam, \( L \) is the antenna feed loss, \( V_r \) is the fraction of the scattering volume which
is filled with scatterers ($V_f = 1$ was assumed) and $A_e$ is the effective area of the receiving antenna. The power received by the radar has been determined as described in Smirnova et al. (2010) using scaling against the daily variation of galactic radio noise. For computation of $\eta$ from Eq. (1), profiles of received power with 2-min time resolution in the 80–90 km height range were used. Volume reflectivity was averaged over 1 h and its maximum over the given altitude range was taken. Values of $L = 1$, $P_1 = 72 \times 10^3$ W, $A_e = 1870$ m² were used, while $A_k$ was changed to 3740 m² after April 2004.

### 2.3 Independent calibration

The antenna loss figure is the most uncertain parameter in calculating the reflectivity. For a new radar, one might expect a loss figure as good as 0.8 (note that $L$ in our case is the one-way loss on transmission as any additional loss on reception is taken into account by the scaling of the detected signal against galactic noise). The value of $L$ for ESRAD was estimated to be 0.39 in August 2006 (Kirkwood et al., 2007). Antenna maintenance work has been done on several occasions since 1997, particularly in 2004 when the radar antenna array was expanded, and in 2008 when major maintenance was undertaken. Thus we expect that $L$ did vary significantly from year to year. $P_1$ may also vary, but since $L$ and $P_1$ appear as a product in Eq. (1), we can account for any small fluctuations in $P_1$ by considering only apparent changes in $L$. There are also a few occasions when there are major dropouts in $P_1$, when whole transmitter blocks fail. Each block consists of 12 kW power and feeds a separate 1/6th of the antenna array, so such dropouts also affect $A_e$ (for transmission). In order to correctly estimate $\eta$, allowing for major dropouts in $P_1$ and possible changes in $L$, we use the result found by Kirkwood et al. (2010a) concerning radar-radiosonde comparisons in the upper troposphere/lower stratosphere (UTLS) region. For the UTLS region, Kirkwood et al. (2010a) showed that Fresnel scatter, as quantified by the normalised power reflection coefficient $\rho^2/\Delta r$, is proportional to $M^2$, where $M$ is the dry term of mean vertical gradient of generalized potential refractive index. The same constant of proportionality was found to apply for 50 MHz radars with very different power-aperture products and in very different locations (ESRAD, the MARA radar in Antarctica and the Indian MST radar at Gadanki). $M$ can be determined from radiosonde measurements according to the formula:

$$M = -77.6 \times 10^{-6} P \frac{\partial \ln \theta}{\partial z},$$

where $P$ is pressure in hPa, $T$ is temperature, $\theta$ is potential temperature, both in $K$.

For any particular radar (e.g. Kirkwood et al., 2010a):

$$\frac{\Delta \rho^2}{\Delta r} = \frac{4 \pi^2 V_f}{64 (2 \ln 2) A_e} \eta.$$

### Table 1. Parameters of ESRAD experiment used in this study.

<table>
<thead>
<tr>
<th>Radar modes</th>
<th>$f_{ca,4500}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pulse length (3 dB)</td>
<td>2.1 $\mu$s (8 bit complementary code with nominal 4 $\mu$s baud)</td>
</tr>
<tr>
<td>Sampling resolution</td>
<td>600 m</td>
</tr>
<tr>
<td>Coherent integrations</td>
<td>128 (98.4 ms) per code up to 2004</td>
</tr>
<tr>
<td>Filter bandwidth</td>
<td>32 (24.6 ms) per code</td>
</tr>
<tr>
<td>Pulser repetition frequency</td>
<td>250 kHz</td>
</tr>
<tr>
<td>Duty cycle</td>
<td>1300 Hz</td>
</tr>
<tr>
<td>Start altitude</td>
<td>4.8 km</td>
</tr>
<tr>
<td>End altitude</td>
<td>105 km</td>
</tr>
<tr>
<td>Receiver filter efficiency $C_{\text{fl}}$</td>
<td>0.72</td>
</tr>
<tr>
<td>Mean altitudes for $(S_{\text{fl}} + S_{\text{sky}})_{\text{nom}}$</td>
<td>30–50 km</td>
</tr>
<tr>
<td>Galactic noise $T_{\text{sky}}$</td>
<td>1680–4500 K</td>
</tr>
</tbody>
</table>

where $\eta$ is calculated using Eq. (1). Hence $\rho^2/\Delta r$ is proportional to $\eta/A_e$ and we can conclude that $\eta/A_e$ must also be proportional to $M^2$, i.e.

$$\eta = B \cdot A_e \cdot M^2,$$

(4)

where $B$ is a constant of proportionality.

First we compute nominal values of volume reflectivity $\eta_{\text{nom}}$ for all heights (covering both UTLS and PMSE echoes) by applying Eq. (1) with $L = L_{\text{coeff}} = 1$. We then find corresponding median values of $B_{\text{nom}}$ for each year from Eq. (4) by comparing $\eta_{\text{nom}}$ with $M^2$ determined from radiosondes.

For calculations of $M^2$, we used data from regular radiosondes launched twice per day in Bodø, Norway, 250 km SW from ESRAD. Radiosondes were launched at 12:00 UT and 24:00 UT, and we used 1-h averaged ESRAD data for the same times. For each ESRAD and radiosonde data set, corresponding to either 12:00 UT or 24:00 UT, we calculated a linear fit of $\eta_{\text{nom}}/A_e$ from ESRAD to $M^2$ derived from the radiosondes, for the altitude range from 8.1 km to 13.5 km, together with the correlation between them. We kept regression coefficients $B$ only when this correlation was statistically significant at the 95% level (for the vast majority of such cases the correlation was higher than 0.5). For these cases, we can assume that the sonde samples the same synoptic air mass as ESRAD. By examining the day-to-day variation of the regression coefficients $B$, including all months of the year, we first identified periods affected by major dropouts in $P_1$ (these affected parts of the PMSE seasons in 2006 and 2007). We then recalculated $\eta_{\text{nom}}/A_e$ taking account of these dropouts and recomputed the corresponding $B$.

Scatter plots of $\eta_{\text{nom}}$ against $M^2$ for June–July data sets for each year are presented in Fig. 1. Note that the scale is logarithmic on both axes. The red lines indicate the radar sensitivity threshold for gradient of potential refractive index.
Fig. 1. Scatter plots of the logarithm of the squared refractive index gradient, $M^2$ derived from radiosonde data against logarithm of radar volume reflectivities, $\eta$ for altitude range from 8.1 km to 13.5 km, for the months of June and July.

(i.e. the level below which there is no correlation). One can see that, for every year, data above the red line lies compactly and with small spread indicating a good correlation between $\eta_{\text{nom}}$ and $M^2$. Below the red line where $M^2$ is relatively small (less than $10^{-17.5}$), there is no useful radar signal, only noise.

The value of $L = 0.39$ evaluated for 2006 is the best available (Kirkwood et al., 2007). It was estimated by comparison of PMSE volume reflectivities between ESRAD and a new, collocated, well calibrated MARA (Movable Atmospheric Radar for Antarctica) radar. However, the loss figure was calculated assuming $P_1 = 72$ kW. Since our comparison with radiosondes has identified a dropout in $P_1$ at the time the comparison was made, a more correct estimate would be $L(2006) = 0.47$. Since the average value of $B$ has been found to be essentially the same in a wide variety of conditions, even at completely different locations over the globe (Kirkwood et al., 2010b), it seems reasonable to assume that it does not vary from year to year at the same location. Thus we can find the appropriate value of $L$ for each year as

$$L(\text{year}) = \frac{B_{\text{nom}}(2006)}{B_{\text{nom}}} \cdot L(2006)$$

Fig. 2. Variations of effective losses in the antenna feed together with their quartiles for 1997–2009.
Fig. 3. The distributions of logarithm of ESRAD volume reflectivity for different years. The histograms refer to maximum $\eta$ over altitudes 80–90 km for one hour averages. White and black bars correspond to noise and PMSE, respectively.

Fig. 4. Year-to-year variations of log$_{10}$ PMSE volume reflectivities (black solid line) together with their quartiles (grey dashed line), K-index (magenta line) and solar flux (blue line).

with $L(2006) = 0.47$, 

$$L(2006) = 0.47,$$  \hspace{1cm} (5)

where $B_{nom}$ are average values over the two summer months, June and July, for each year. The resultant values of $L$ are shown in Fig. 2. From this figure it is clear that $L$ has had high values close to 0.8–0.9 in the first few years of operation (up to year 2000) and from 2008 onwards. Between summer 2000 and 2001 the initial stages of work to extend the antenna array to twice the area were started and some reduction in $L$ is noticeable. The full enlarged antenna was connected to the system in April 2004, leading to further degradation in $L$, which subsequently varied from year to year as minor maintenance was carried out. Finally, following major maintenance in 2008, $L$ returned to a relatively high value.

For the PMSE, we can correct our $\eta_{nom}$ for the variation of $L$

$$\eta(\text{year}) = \frac{\eta_{nom}}{L(\text{year})}$$  \hspace{1cm} (6)

For future comparison with results from other radars it can be noted that, for the UTLS echoes, the corresponding average value of $(\rho^2/\Delta r)/M^2 = 1.1 \times 10^{-3}$. 

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3 Results

3.1 Distributions of volume reflectivity strengths

We have calculated the loss-corrected maximum ESRAD volume reflectivities over the altitude range 80–90 km, with 1 h averaging, for the interval from 1 June to 31 July for each year. The histogram of PMSE reflectivity strengths is presented in Fig. 3. PMSE signals were defined to correspond to reflectivities exceeding the threshold value of $10^{-15}$ m$^{-1}$ and are shown by black bars. For consistency we have used the same threshold as used for calculation of PMSE OR in Smirnova et al. (2010). This value is well above the background noise (white bars).

In general the distributions for each year are a superposition of two populations, PMSE and noise. The location of the maximum in the volume reflectivity distribution varies from $2 \times 10^{-15}$ to $3 \times 10^{-14}$ m$^{-1}$. In 2002 and even more so in 2007, the most probable PMSE strength was very low, close to the chosen threshold. (It can also be noted that the distributions of PMSE strength are far from Gaussian so that mean values are of limited use.)

3.2 Variation of volume reflectivity over 1997–2009 and its relation to solar and geomagnetic activity

Bremer et al. (2006, 2009) have shown that year-to-year variations of PMSE OR are to a large extent correlated with geomagnetic and solar activity. Smirnova et al. (2010) have also found a significant positive correlation between PMSE OR and geomagnetic activity. We have tested this hypothesis for PMSE volume reflectivity. A proxy of solar activity is the 10.7 cm radio flux, while the K-index represents the local geomagnetic activity during 3-h intervals. We use local K-index rather than planetary k- or a-index because we want to study the relationship between $\eta$ and geomagnetic activity at the same location during different times of the day.

We calculated seasonally averaged PMSE volume reflectivities according to the following procedure. Firstly, we calculated the daily means of loss-corrected ESRAD volume reflectivities averaged over 1 h and exceeding the chosen threshold of $10^{-15}$ m$^{-1}$. And then we took a median over the time interval from 1 June to 31 July. Mean values for the same time interval for 10.7 cm solar flux and daily averaged K-index were also calculated. In Fig. 4 the behaviours of all three quantities over the years 1997–2009 are presented. One can easily notice that radar $\eta$ changes closely follow those in K-index at least until 2006. However, any solar cycle signature in $\eta$ is, if present, not very pronounced. In order to quantify the relation between these three parameters, and allow for a possible long-term trend, we have calculated the multi-parameter linear regression as follows:

$$\eta = A_0 + A_1 \times (\text{year} - 1997) + A_2 \times (F - F_{\text{mean}}) + A_3 \times (K - K_{\text{mean}}).$$

(7)

We have calculated the same “stepwise” function allowing finding the best fit by interactively moving predictors in and out of the fit in accordance with their statistical significance. The method begins with an initial model having no terms and holds a few steps. At each step the level of significance for F-statistic (p-values) is computed to test models with and without a potential term. We test p-values for all possible combinations of terms and retain those terms which give p-values less than 0.05.

The results of the final fits are presented in Table 2. It shows that the year-to-year variation of PMSE $\eta$ and geomagnetic index are positively correlated throughout the entire day. The correlation between PMSE $\eta$ and solar flux is always negative and significant for the interval 00:00–09:00 UT. For daily mean data the radar volume reflectivity is statistically significantly correlated with solar flux (negatively) and with magnetic index (positively). The only significant positive trends in the seasonal median behaviour of $\eta$ are found for the interval from 09:00 to 12:00 UT. There is no trend for daily mean data. Table 2 also presents the coefficients of determination ($R^2$) which, for a linear regression, are equal to the square of the correlation coefficient between observed and modelled (linear fit) data values. $R^2$ determines the proportion of variability in $\eta$ that is accounted for by two predictors: F10.7 cm solar flux and K-index. We see that, for example, for daily averaged data 86% of the year-to-year variation of PMSE volume reflectivity can be explained by solar and geomagnetic activity.

It is instructive to compare these results, based on PMSE volume reflectivities, to previously published results based on PMSE occurrence rate. For example, Smirnova et al. (2010) analysed OR using measurements from 1997–2008, also from ESRAD. While the range of year-to-year variations in $\eta$ is more than an order of magnitude, the range in OR is only $\pm 15\%$. So $\eta$ should be a more sensitive indicator of year-to-year changes. It is notable that the signs of the correlations between OR and magnetic activity (positive) and solar flux (negative) were found to be the same as now found for $\eta$, but the statistical certainty of the result is much higher using $\eta$ than OR (the p-value of the final fit was 0.11 for OR, and 0.0001 for $\eta$). Bremer et al. (2009) used measurements from 1994–2008, made by radars located $\sim 200$ km north-west of ESRAD. They found a similar range in OR as at ESRAD, a similar positive correlation with magnetic activity, but also a positive correlation with solar activity. However
Table 2. The coefficients of multi-parameter regression for year-to-year variation of PMSE $\eta$ together with their 95% confidence intervals and $p$-values of F-statistics for testing hypothesis of a zero coefficient. $P$ is the $p$-value of F-statistic for testing the hypothesis of all zero coefficients and $R^2$ is the square of the correlation coefficient between original $\eta$ and the final fit. The parentheses indicate that coefficient was set to zero in the final fit.

<table>
<thead>
<tr>
<th>UT (h)</th>
<th>$A_0$ (10$^{-14}$)</th>
<th>$A_1$ (10$^{-15}$)</th>
<th>$A_2$ (10$^{-16}$)</th>
<th>$A_3$ (10$^{-13}$)</th>
<th>$P$</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>00:00-03:00</td>
<td>4.9 (4.1)</td>
<td>$-12.5 \pm 9.1$</td>
<td>$8.6 \pm 2.8$</td>
<td>0.0008</td>
<td>0.79</td>
<td></td>
</tr>
<tr>
<td>03:00-06:00</td>
<td>6.3 (2.1)</td>
<td>$-21.5 \pm 14.3$</td>
<td>$18.8 \pm 6.7$</td>
<td>0.0004</td>
<td>0.82</td>
<td></td>
</tr>
<tr>
<td>06:00-09:00</td>
<td>2.5 (1.3)</td>
<td>$-4.2 \pm 1.9$</td>
<td>$4.6 \pm 1.1$</td>
<td>0.0002</td>
<td>0.91</td>
<td></td>
</tr>
<tr>
<td>09:00-12:00</td>
<td>1.2 (0.3)</td>
<td>$-2.0$</td>
<td>$4.0 \pm 1.5$</td>
<td>0.0006</td>
<td>0.80</td>
<td></td>
</tr>
<tr>
<td>12:00-15:00</td>
<td>3.2 (1.4)</td>
<td>$-4.8$</td>
<td>$2.4 \pm 1.9$</td>
<td>0.02</td>
<td>0.43</td>
<td></td>
</tr>
<tr>
<td>15:00-18:00</td>
<td>1.6 (1.1)</td>
<td>$-2.2 \pm 2.0$</td>
<td>$1.9 \pm 1.2$</td>
<td>0.02</td>
<td>0.60</td>
<td></td>
</tr>
<tr>
<td>18:00-21:00</td>
<td>1.1 (1.3)</td>
<td>$1.7 \pm 0.8$</td>
<td>$0.7 \pm 0.6$</td>
<td>0.02</td>
<td>0.45</td>
<td></td>
</tr>
<tr>
<td>21:00-24:00</td>
<td>2.6 (1.6)</td>
<td>$-4.2$</td>
<td>$2.8 \pm 1.7$</td>
<td>0.005</td>
<td>0.56</td>
<td></td>
</tr>
<tr>
<td>00:00-24:00</td>
<td>7.0 (5.8)</td>
<td>$-16.7 \pm 8.9$</td>
<td>$1.8 \pm 0.6$</td>
<td>0.0001</td>
<td>0.86</td>
<td></td>
</tr>
</tbody>
</table>

their data set suffers the disadvantage that two quite different radars were used before and after 1998, introducing considerable uncertainty in the relative calibration. Also, the $p$-value of their result for the dependence of OR on solar activity ($>0.1$) was lower than the present result for $\eta$ ($<0.003$). (Both estimates based on the same Fisher F-test.)

4 Discussion

4.1 PMSE volume reflectivity and solar flux

Variations of solar flux can influence PMSE strength in at least three different ways: via changing ionization, the amount of water vapour and the atmospheric temperature in the upper mesosphere and lower thermosphere region.

Electron density in this region is mainly determined by NO ionization by the solar Lyman $\alpha$ flux which is expected to vary in the same way as the solar 10.7 cm radio flux, with values ranging from 3.5 to $5.5 \times 10^{11}$ photons cm$^{-2}$ s$^{-1}$, between minimum and maximum solar activity during the interval from 1997 to 2009 (Woods, 2008, see also Fig. 7 in Bremer et al., 2006). An increase in solar Lyman $\alpha$ radiation leads to increased water vapour photo-dissociation (Brasseur and Solomon, 1986). This should result in less ice-particle formation (von Zahn et al., 2004) and hence should affect PMSE. Variation of solar radiation could also influence atmospheric temperature via production of ozone after photodissociation of O$_2$ (Robert et al., 2010) and via changing atmospheric dynamics.

Kirkwood et al. (2010b) have found an empirical relation between PMSE $\eta$ and ionospheric plasma and ice particle parameters using the 54 MHz radar MARA at Wasa, Antarctica and the SOFIE instrument onboard the AIM satellite. Radar volume reflectivity was found to be proportional to the product of background electron density and ice mass density. The result was based on PMSE observations during one season (for a time interval close to local midnight) and restricted to geomagnetically quiet conditions with $Ap < 15$, when there was no correlation between PMSE $\eta$ and $Ap$. If we apply this empirical relation, then for increased solar flux the enhanced electron density will counteract the decrease of ice mass due to depletion of water vapour and increased temperature. The net effect on $\eta$ depends on which factor prevails. Thus $\eta$ might either correlate or anti-correlate with solar flux. From our regression (Table 2) we can conclude that the net effect is negative. Similarly, anti-correlations with the solar cycle were found for both polar mesospheric cloud (PMC) albedo and occurrence (with 0.5–1 year shift) using satellite-borne observations covering 27 years (DeLand et al., 2007; Shettle et al., 2009). NLC occurrences from ground-based
The electric field of those in PMSE (to 06:00 UT, when the correlation between each 3-h time interval together with probability (η) present correlation coefficient of electric field) and hence electron density, and electrojet currents which depend on both conductance (height-component of the geomagnetic field). Kellerman et al. (2009) showed that the average number of high background electron density, PMSE strength is controlled by ice density $N_d$. Our result does not contradict that obtained by Varney et al. (2010) because the analytical expression for η obtained there cannot be directly applied to the year-to-year variations shown in Fig. 4. One can use that formula for qualitative analysis for two extreme cases when $N_e \ll N_d$ or $N_e \gg N_d$. Hervig et al. (2009) reported that the average $N_d$ is $2\times10^8$ m$^{-3}$ but can reach $10^9$ m$^{-3}$. Electron densities at 80–90 km altitudes over Kiruna in summer time are determined by ionization by solar UV flux and by energetic magnetospheric particles which, in turn, vary with geomagnetic activity (Codrescu et al., 1997). For nighttime and geomagnetically quiet conditions night-time only the geomagnetic field dominates during daytime. In Table 3 we present correlation coefficients between η and K-index for each 3-h time interval together with probability (p) that there is no correlation between them. We add to this table the results from Fig. 6 of Kellerman et al. (2009) for the correlation between perturbations in the horizontal component of the geomagnetic field ($\delta H$), which is the basis of the K-index, cosmic noise absorption ($A$) representing conductance due to energetic particle precipitation, and ion convection velocity ($V_i$) as a proxy for ionospheric electric field. From 18:00 UT to 06:00 UT, when the correlation between $\delta H$ and $A$ is high ($>0.5$), it is reasonable to suggest that electron density variations due to energetic particle precipitations may be the cause of those in PMSE η. In the daytime (06:00 UT–15:00 UT) the electric field contributes most to K-index perturbations and hence perhaps also to those for η. The mechanism of such an influence is not obvious although vertical transport, as for metallic-ion sporadic-E layers might be involved (see e.g. Kirkwood and Nilsson, 2000). This possibility needs to be studied further.

### Table 3. Correlation coefficients.

<table>
<thead>
<tr>
<th>Time, UT</th>
<th>Corr (η, K)</th>
<th>Time, UT</th>
<th>Corr (δH, A)*</th>
<th>Corr (δH, $V_i$)*</th>
</tr>
</thead>
<tbody>
<tr>
<td>UT = SLT – 1.5 h</td>
<td></td>
<td>UT = MLT – 2.5 h</td>
<td></td>
<td></td>
</tr>
<tr>
<td>00:00–03:00</td>
<td>0.76</td>
<td>00:30–03:30</td>
<td>0.65</td>
<td>0.52</td>
</tr>
<tr>
<td>03:00–06:00</td>
<td>0.77</td>
<td>03:30–06:30</td>
<td>0.50</td>
<td>0.27</td>
</tr>
<tr>
<td>06:00–09:00</td>
<td>0.82</td>
<td>06:30–09:30</td>
<td>0.08</td>
<td>0.56</td>
</tr>
<tr>
<td>09:00–12:00</td>
<td>0.82</td>
<td>09:30–12:30</td>
<td>0.17</td>
<td>0.90</td>
</tr>
<tr>
<td>12:00–15:00</td>
<td>0.66</td>
<td>12:30–15:30</td>
<td>0.09</td>
<td>0.72</td>
</tr>
<tr>
<td>15:00–18:00</td>
<td>0.59</td>
<td>15:30–18:30</td>
<td>0.30</td>
<td>0.59</td>
</tr>
<tr>
<td>18:00–21:00</td>
<td>0.67</td>
<td>18:30–21:30</td>
<td>0.69</td>
<td>0.28</td>
</tr>
<tr>
<td>21:00–24:00</td>
<td>0.75</td>
<td>21:30–00:30</td>
<td>0.76</td>
<td>0.38</td>
</tr>
</tbody>
</table>

* Correlation coefficients were adopted from Fig. 6a of Kellerman et al. (2009) and UT time was recalculated from magnetic local time (MLT) time for Kiruna as UT = MLT − 1.5 h. $p$ is probability that there is no correlation between η and K-index.
Thus we averaged $\eta$ when electron density was high and low which makes the formula for $\eta$ by Varney et al. (2010) inapplicable for our result.

### 4.3 Long-term trend

For one 3-h interval (09:00–12:00 UT), a statistically significant positive trend of $2.5 \times 10^{-15} \text{m}^{-1} \text{year}^{-1}$ was found. This gives an increase of PMSE volume reflectivities by $3.3 \times 10^{-14} \text{m}^{-1}$ over 13 years, which is comparable to the mean value of $\eta$ over this interval, $2.9 \times 10^{-14} \text{m}^{-1}$. However the regression coefficient for the variation of $\eta$ with the K-index is $4 \times 10^{-11} \text{m}^{-1}/K$ and $K$ varies from a minimum value of 1.5 to a maximum value of 4. Thus K-index variations rather then a long-term trend primarily determine the behaviour of PMSE strength during the 13 years.

For daily averaged data we did not find any statistically significant trends in PMSE absolute strength during 1997–2009. As already mentioned in the introduction, we have a unique PMSE data set from which we were able to calculate seasonal means of PMSE $\eta$ including correction for changing antenna losses and transmitted power. We cannot compare our results with others since, to our knowledge, so far there is no such study regarding PMSE strength. However, there are a number of publications related either to PMSE occurrence rate (OR) or to NLC/PMC OR. Both Bremer et al. (2009) and Smirnova et al. (2010) did not find statistically significant trends in PMSE OR during 1994–2008 and 1997–2009, respectively. Similarly, Kirkwood et al. (2008) reported an absence of statistically confident trends in OR of moderate or bright NLCs over 43 years of observations. In contrast, positive and significant trends of up to 20% in OR of PMCs have been found in observations by satellites borne instruments over the last 27 years (Shettle et al., 2009). This is too small to be visible on the shorter interval of 13–15 years of PMSE observations and on the background of the large inter-annual variability. Moreover, PMSE OR is a complex quantity where trends/variabilities will depend on the interplay between those in water vapour and temperature (as for NLC/PMC) as well as in ionospheric conditions. Some studies showed indications of positive trends in the last 20–50 years in water vapour concentration (e.g. Lübken et al., 2009) and electron density below 90 km (e.g. Lastovicka and Bremer, 2004). Thus one could expect positive trends in PMSE strength and OR also. However, their values maybe not be large enough to be visible on an interval of just 13–15 years.

### 5 Summary and outlook

We have calculated the radar volume reflectivities for PMSE observed over Esrange, Northern Sweden during summers 1997–2009.

For the first time we have presented a new independent calibration method which allows estimation of the possible losses in antenna feed (and transmitter power) over the years and improves the accuracy of calculations for radar volume reflectivity.

The distribution of $\eta$ over magnitudes (histogram) varies from year to year with distribution maximum (peak of the histogram) lying in the range from $2 \times 10^{-15}$ to $3 \times 10^{-14} \text{m}^{-1}$.

Year-to-year variations of $\eta$ were found to be strongly correlated with the local geomagnetic K-index as averaged over 24 h, and for every 3-h interval. One possible explanations of the positive correlation is that increased geomagnetic activity can lead to enhanced ionization at altitudes 80–90 km and hence to enhanced $\eta$. There is also a hint that ionospheric electric field might influence PMSE strength, however this requires additional study.

In order to test the statistical significance of trend, solar cycle variation and geomagnetic activity in the year-to-year variability of PMSE strength, we have made a multi-parameter linear fit to PMSE $\eta$, which includes all of these three factors. We found that both solar 10.7 cm flux and K-index can explain 86% of inter-annual variations of $\eta$ averaged over 24 h, where solar flux anti-correlates and K-index correlates with $\eta$. No statistically significant trends in PMSE yearly strengths were found over an interval of 13 years.

The conclusion regarding a possible relation between PMSE strength and ionospheric electric field discussed in Sect. 4.2 is not based on direct measurements of electric field and needs further study. This could be accomplished using e.g. the European incoherent scatter (EISCAT) VHF and UHF radars for simultaneous measurements of both quantities.

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**References**


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Abstract. Aspect sensitivities of polar mesosphere summer echoes (PMSE) measured with the ESRAD 50 MHz radar in 1997-2010 are calculated using the full correlation analysis technique. Half of the PMSE detected each year prove to be highly aspect sensitive. Yearly median values of aspect sensitivity parameter $\theta_s$, characterising half-width of the scatterers' polar diagram, are 2.9-3.7° depending on the year. Another half of PMSE have $\theta_s$ values larger than 9-11° and cannot be evaluated using the ESRAD vertical beam only. PMSE aspect sensitivity reveals an altitude dependence, the scatter becomes more isotropic with increasing height. This result is consistent with that reported in the other earlier studies. No dependence of PMSE aspect sensitivity on backscattered power for any year was identified. The paper discusses the limitations of the in-beam and off-vertical beam methods for estimation of PMSE aspect sensitivity. We conclude that both methods should be combined in order to get complete information about PMSE aspect sensitivity and to estimate correctly PMSE absolute strength.
1. Introduction

Polar Mesosphere Summer Echoes (PMSE) are extremely strong radar echoes observed from altitudes of 80-90 km at high latitudes in the Northern and Southern Hemispheres during summer. Nowadays it is known that PMSE occur due to scattering from fluctuations in electron density caused by atmospheric turbulence in the presence of ice particles formed from water vapour during low temperatures (~130 K) at the summer mesopause. Since PMSE are closely related to temperature changes and probably follow climate change, they have attracted considerable interest from the scientific community over the past three decades. A review of PMSE investigations can be found in Rapp and Lübken (2004).

Despite long and intensive studies of PMSE, their aspect sensitivity is still an intriguing and open question. Aspect sensitivity is a property of the scatterers and describes the variation of scattered power in respect of incident angle. It is quantified in terms of the half width $\theta_s$ of the angular polar diagram of backscatter (Hocking et al., 1986). Isotropic scatterers are non-aspect sensitive and have a broad polar diagram, while anisotropic, specular ones are highly aspect sensitive, i.e. have a narrow backscatter beam. The early measurements of aspect sensitivity of PMSE by Czechowsky et al. (1988) showed that they are highly aspect sensitive with, typically, $\theta_s$ of 5°-6°. Further work on this subject reported a wide variety of $\theta_s$: e.g. 12°-13° (Huaman and Balsley, 1998), 3.5° and 7°-10° (Zecha et al. 2001), 8°-13° (Swarnalingam et al., 2011). This implies that PMSE can be specular as well as rather isotropic.

Why is aspect sensitivity interesting for PMSE researchers? Firstly, as we shortly discussed above, aspect sensitivity measurements can provide some idea about scattering processes and the scatterers themselves. Swarnalingam et al. (2011) gave a short overview of the main models of coherent scatter for VHF radars. Not all radar observations in the middle atmosphere were consistent with the two classical extreme models: turbulent volume isotropic scatter and specular Fresnel reflection. Therefore new models for anisotropic turbulence and Fresnel scatter were suggested (for references see Swarnalingam et al., 2011).

There is another important effect of PMSE aspect sensitivity which has not received proper attention yet. Knowledge of aspect sensitivity could be essential for correct estimation of the absolute strength of PMSE. For instance, highly aspect
sensitive echoes do not fill a whole radar sampling volume. Then with the usual assumptions of the volume (isotropic) scatter of 100% filling, one could underestimate the actual volume reflectivity of PMSE. Another possible application area for the aspect sensitivity measurements is in inter-comparisons of PMSE observed with different radars. These studies, including PMSE interhemispheric and latitudinal differences have become very popular in recent years (Kirkwood et al, 2007; Latteck et al., 2008; Morris et al, 2009; Swarnalingam et al., 2009a). However, corrections should be made in the calculations of PMSE strength because the radars involved in these studies have various beam-width antennas and PMSE at the radar locations might have different aspect characteristics. Moreover, isotropic (turbulent) and Fresnel scattering mechanisms have very different cross-sections (e.g. Kirkwood et al., 2010). Therefore in the cases of isotropic and aspect-sensitive PMSE, estimation of their cross-sections using observations requires quite different calculations to account for radar characteristics. For a given echo power at the radar receiver, volume reflectivity depends on the antenna effective area, whereas Fresnel reflectivity depends on the square of that area.

Aspect-angle sensitivities have been obtained from radar measurements by two different methods: (1) by comparing the echo strengths from vertical and off-vertical radar beams, or, in another terminology, Doppler beam swinging, DBS; (2) by in-beam estimates using spaced antennas (SA). The latter applies coherent radar image (CRI) or the full correlation analysis (FCA) techniques. Hobbs et al. (2001) described both methods and their limitations resulting from the theoretical assumptions and experimental configurations. Chilson et al. (2002) pointed out that in an application to the mesosphere a disadvantage of the DBS method is in the long distance between the radar sampling volumes because large off-zenith angles are used. We will discuss more limitations for both methods later in this paper.

In the majority of PMSE aspect sensitivity studies the DBS method has been used (Czechowsky et al., 1988; Hoppe et al, 1990; Huaman and Balsley, 1998; Swarnalingam et al., 2011 and others). Chilson et al. (2002) applied CRI technique to PMSE and related the angular brightness distribution to aspect sensitivity. Zecha et al. (2001) used both DBS and FCA methods for measuring PMSE aspect angles and obtained results which significantly differ from each other. The authors argued that these two methods evaluate aspect sensitivity at different spatial scales and PMSE is strongly anisotropic within a radar beam but more isotropic over larger distances.
There are also several papers dealing with indirect estimation of aspect sensitivity of the mesospheric summer echoes, e.g. via the relationship between the echo power and its spectral width (e.g. Chen et al, 2004). However, we will not touch on those results in this paper where we concentrate on quantitative characterisation of aspect sensitivity, i.e. on calculation of $\theta_s$.

Here we present the results of PMSE aspect sensitivity measurements over the period 1997-2010 using the ESRAD MST radar located near Kiruna in Northern Sweden. We have used the FCA technique which provides us with in-beam estimates of aspect angle $\theta_s$. We discuss a limitation of this technique in application to the ESRAD radar. We analyze the dependence of aspect sensitivity on PMSE height as well as on backscattered power and compare our results with those of others. Finally, we evaluate the effect of PMSE aspect sensitivity on estimation of their volume reflectivity.

2. Experiment description

PMSE measurements have been carried out with the ESRAD 52 MHz radar, situated at the rocket range Esrange, Sweden (67.88° N, 21.10° E), during the years 1997-2010, although observations from 1999 are not considered here due to a radar malfunction. ESRAD provides information on the dynamic state of the lower and middle atmosphere such as winds, waves, turbulence and layering. A detailed description for ESRAD was given by Chilson et al. (1999) and updated by Kirkwood et al. (2007).

For PMSE measurements ESRAD was operated in different modes with the radar beam pointed vertically. In this paper we consider the data with the highest available altitude resolution of 150 m. The measurements collected in 2002 and 2004 are not included since, during these years, ESRAD was operated in modes providing only 300 m or 600 m altitude resolution. Initially the $e^{-1}$ half-width of the radar transmit beam was 3.98°, then in 2004 the antenna array was extended, making this radar beam width narrower, 2.65°. The details of the radar measurement set-up are presented in Table 1.

ESRAD was operated in 6-receiver mode which allows the implementation of FCA analysis. The FCA technique was developed by Briggs (1985). The principle behind this technique is the following: the scatterers lead to a diffraction pattern on the
ground, which moves across the antenna array as the scatterers drift horizontally. By using several (at least three) non-collinear antenna sub-arrays for receiving and calculating cross-correlations between them, under certain assumptions one can derive spatial parameters of the diffraction pattern. Finally, from them the aspect sensitivity is estimated. We consider the calculation of aspect sensitivity in more detail in the next section.

Table 1. ESRAD operating parameters/modes used in this study

<table>
<thead>
<tr>
<th>Radar parameter/mode</th>
<th>fca_150 for 1997-2002</th>
<th>fca_150 for 2003-2010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transmitter peak power</td>
<td>72 kW</td>
<td>72 kW</td>
</tr>
<tr>
<td>e-1 half-beam width</td>
<td>3.98°</td>
<td>2.65°</td>
</tr>
<tr>
<td>Pulse length (3 dB)</td>
<td>1 μs</td>
<td>1 μs</td>
</tr>
<tr>
<td>Sampling resolution</td>
<td>150 m</td>
<td>150 m</td>
</tr>
<tr>
<td>Code</td>
<td>8/16-bit complementary</td>
<td>none</td>
</tr>
<tr>
<td>Coherent integrations</td>
<td>64/128</td>
<td>256</td>
</tr>
<tr>
<td>Pulse repetition frequency</td>
<td>1450 Hz</td>
<td>4688 Hz</td>
</tr>
</tbody>
</table>

3. Calculation of aspect sensitivity

A radar transmits radiation in a certain beam pattern, which can be expressed by a polar diagram. When radiation reaches scatterers they reradiate back with a polar diagram of backscatter defined by their aspect sensitivity. This backscatter is received by the radar in accordance of its receive beam which can differ from the transmit one (as it does for the FCA technique). The backscatter pattern that we receive back at the radar can be described by an effective polar diagram. We follow Hocking (1986) and approximate every polar diagram by the expression:

\[ P(\theta) = \exp \left( -\frac{\sin^2 \theta}{\sin^2 \theta_0} \right), \]  

(1)

where \( \theta \) is the zenith angle and \( \theta_0 \) is the e-1 half-width of the polar diagram. Then the effective polar diagram is the product of polar diagrams of transmit and receive beams and that of the scatterers, i.e.

\[ P_{\text{eff}}(\theta) = \exp \left( -\frac{\sin^2 \theta}{\sin^2 \theta_{\text{eff}}} \right) = \exp \left[ -\frac{\sin^2 \theta}{\sin^2 \theta_T} - \frac{\sin^2 \theta}{\sin^2 \theta_R} - \frac{\sin^2 \theta}{\sin^2 \theta_S} \right], \]  

(2)

where \( \theta_{\text{eff}} \) is the e-1 half-width of the effective polar diagram, \( \theta_T \) is the e-1 half-width of the radar transmit beam, \( \theta_R \) is the e-1 half-width of the radar received beam, \( \theta_S \) is the e-1 half-width of the scatter polar diagram of the scatterers. \( \theta_S \) is a measure of aspect
sensitivity, or in the same terminology as e.g. in Swarnalingam et al. (2011), it is the aspect sensitivity parameter.

We made use of FCA results for the calculation of $\theta_{\text{eff}}$ as described by e.g. Holdsworth (1995):

$$
\theta_{\text{eff}} = \frac{15.2 \lambda \sqrt{R_{\text{ax}}}}{S_{0.5}},
$$

(3)

where $\lambda$ is the radar wavelength and $R_{\text{ax}}$ and $S_{0.5}$ are the FCA estimated axis ratio and scale of the diffraction pattern, respectively, estimated by the FCA technique.

Finally, for a given $\theta_T$ and $\theta_R$ the aspect sensitivity parameter $\theta_S$ can be found from the equation:

$$
\sin^{-2} \theta_{\text{s}} = \sin^{-2} \theta_{\text{eff}} - (\sin^{-2} \theta_T + \sin^{-2} \theta_R).
$$

(4)

Finite transmit and receive radar beam widths set limits for in-beam measurements of aspect sensitivity. If, for instance, a radar beam is narrow, one cannot distinguish between two types of scatterers whose backscatter beam widths (i.e. aspect sensitivity) are larger than that of the radar. In order to study this with respect to ESRAD we computed $\theta_{\text{eff}}$ using eq (4) for various values of aspect sensitivity for the actual ESRAD transmitted beam width. Six rectangular antenna subarrays were used for the reception (each 6 x 8 yagi antennas, 4 x 6 before 2004), with mean beam width $\theta_R$ 6° (9° before 2004) as described in more detail by Kirkwood et al. (2010). In Fig.1, we present the effective polar diagram width $\theta_{\text{eff}}$ as a function of aspect sensitivity parameter $\theta_S$.

![Fig. 1](image_url)

Fig. 1. The $e^{-1}$ half-width $\theta_{\text{eff}}$ of the effective polar diagram of received signal as a function of aspect sensitivity parameter $\theta_S$ (the $e^{-1}$ half-width of the scatter polar diagram of the scatterers) for two radar half-beam widths: 3.98° before 2004 and 2.65° after 2004.
From this figure we see that $\theta_{\text{eff}}$ depends strongly on $\theta_s$ only for the interval from 0° up to about 3.4° (2.3° after 2004). This gives us maximum values of $\theta_{\text{eff}}$ which can be used to derive of aspect sensitivity from the ESRAD data and maximum $\theta_s$ of about 11° (9° after 2004) which we are able to detect using the ESRAD vertical beam.

4. Results

4.1 Aspect sensitivity of PMSE

Firstly, we computed $\theta_{\text{eff}}$ using FCA analysed data for the altitude range 80-90 km for the interval from 1 June to 31 July for each year. FCA is applied only for echoes with signal-to-noise ratio more than 1 db. Then by applying limits discussed in the previous section we selected $\theta_{\text{eff}} < 3.4°$ (2.3° after 2004) for further calculations of the aspect sensitivity parameter.

Fig. 2 shows one typical example of PMSE aspect sensitivity measurements during one day together with the radar backscattered power. It is evident from the figure that

Fig. 2. Backscattered power (the upper panel) and aspect sensitivity (the lower panel) measured with ESRAD MST radar on 3 July 2010.
radar echoes are highly aspect sensitive (with $\theta_s$ less than 2.2°) at the edges of PMSE layers, especially at the lower edges, and they are less aspect sensitive in the middle of the layers. The same features were observed during other days.

In order to study the variability of aspect sensitivity of PMSE from year to year we calculated distributions of aspect sensitivity parameter for each year. These are shown in Fig. 3. The number of data points used for calculation of these distributions varies from year to year mainly because of changes in time allocation for the different radar observation modes. The distributions are very similar, non-Gaussian in shape and have a peak value of about 2.5° and half maximum full width of 2°-3°. The median values of $\theta_s$ are in the range of 2.9°-3.7°. This implies that PMSE are rather aspect sensitive. We calculated also the percentage of such echoes ($\theta_{eff} < 3.4°$ or 2.3° after 2004) in the whole data set for each year. The results show that 50% - 52% of data are represented by the distributions in Fig.3, i.e. half of all PMSE observed with ESRAD are highly aspect sensitive. The other half of PMSE have aspect sensitivity parameters larger than 11° (9° after 2004), i.e. scatter is more isotropic. However, we cannot calculate their exact values from the ESRAD data using the in-beam method.

Fig. 3. The distributions of aspect sensitivity parameter for different years. N is a number of data points, MAS is a median aspect sensitivity parameter in degrees.
4.2 Dependence of PMSE aspect sensitivity on altitude

To check the dependence of PMSE aspect sensitivity on altitude, we computed the distribution of aspect sensitivity parameter $\theta_s$ for each 150-m altitude range for every year. In Fig. 4 we present the results for 1998 for the initial antenna array configuration and for 2010 for the extended antenna array. There each row represents an aspect sensitivity distribution function at a certain height, with values (fraction of the data or occurrence frequency) depicted by colours. The pictures for 1998 and 2010 look very similar. The same behaviour of the PMSE aspect sensitivity is seen for the other years (not shown).

![Fig. 4](image1)

**Fig. 4.** The first and third panels: distribution of aspect sensitivity parameter for each 150-m altitude interval for 1998 and 2010, respectively. Each row is normalized so that the sum of all data bins in it is 1. The second and fourth panels: number $N$ of data points contributing to each altitude interval for 1998 and 2010, respectively.

![Fig. 5](image2)

**Fig. 5.** Left panel: distribution of aspect sensitivity parameter for each 150-m altitude interval averaged over all 11 years. Each row is normalized so that the sum of all data bins in it is 1. Right panel: number $N$ of data points contributing to each altitude interval.
Fig. 5 shows the $\theta_s$ altitude-frequency plot averaged over all 11 years. With increasing altitude the distribution becomes broader and the peak of the distribution moves to higher values of $\theta_s$. We calculated that PMSE median aspect angle for the seasons 1997-2010 at 87-88 km altitude is 1-2º larger than that at 81-82 km. Thus we can conclude that PMSE are more isotropic at the higher altitudes.

### 4.3 Dependence of PMSE aspect sensitivity on power

We have investigated the dependence of PMSE aspect sensitivity on backscattered power for each PMSE season. We computed distribution functions of the aspect sensitivity parameter $\theta_s$ for each 0.2 step in the logarithm of the echo power. We present the results for 1998 and 2010 in Fig. 6. In a similar way to Fig. 4, each row here represents an aspect sensitivity distribution function at certain value of PMSE backscattered power. Again we see that distributions for 1998 and 2010 are very similar and do not show an obvious dependence of the PMSE aspect sensitivity on echo strength. (For the high power there is an apparent decrease of aspect sensitivity, however this result is based on the poor statistics – see the right panel). Fig. 7 shows the distributions averaged over all 11 years, which reveals that PMSE aspect sensitivity is independent of echo backscattered power.
5 Discussion

As mentioned in the introduction, the aspect sensitivity of PMSE has been measured using two methods: in-beam as in this paper and with tilted beams. Both methods suffer from limitations, some of them were considered by e.g. Hobbs et al. (2001) and Chilson et al. (2002). Here we discuss limitations related only to a finite radar beam-width. In section 3 we obtained the largest aspect angles $\theta_s$ which are measurable with ESRAD using FCA. They are 9-11° and determined by the two-way radar beam width. Thus using the ESRAD vertical beam we cannot measure isotropic turbulent echoes. In contrast, for measurements using tilted beams the off-zenith angle should be at least one full radar beam-width (two-way) or larger so that the sampling volumes for the vertical and tilted beams do not overlap each other. Hence the titled beam method is not suitable for quasi-specular echoes whose angular polar diagram width is less than the radar beam-width. This implies that by using one of the two techniques we restrict ourselves to measurements of the part of the echoes produced by a certain type of scatterers. The values of aspect sensitivity parameter obtained with these two methods support our statement. Indeed, in-beam measurements of PMSE aspect sensitivity by Zecha et al. (2001) using FCA and by Chilson et al. (2002) using CRI, resulted in typical values of $\theta_s$ of 2-5° and 2-3°, respectively. We found the most typical $\theta_s$ to be in range from 1° to 5° which agrees well with those found by Zecha et al. and Chilson et al. In turn, using the 8° off-zenith beam, Zecha
et al. (2001) obtained $\theta_s$ in the range mostly 4-11°. Huaman and Balsley (1998) reported $\theta_s$ of 12-13° when a 15° off-vertical beam was used. Most recently, Swarnalingam et al. (2011) found PMSE aspect angles of 8-13° using tilted beams at 10.9° from zenith. The only result deviating from these values is an estimate that $\theta_s$ ranged from 2° to 10° reported by Czechowsky et al. (1988). A possible explanation is that they used beams tilted only 4° and 5.6° off-vertical and the radar had a very narrow full beam-width of 3° (one-way). Thus we can conclude that the in-beam method allows detection of small aspect angles (just a few degrees) and correspondingly highly aspect sensitive echoes, whereas the method using off-vertical radar beams detects less aspect sensitive echoes.

On the basis of 11 years of observations we found that PMSE are more isotropic, less aspect sensitive, at higher altitudes. A similar height dependence of PMSE aspect sensitivity has been reported in other studies (Czechowsky et al., 1988; Zecha et al., 2001, Chilson et al., 2002). Long-term aspect sensitivity measurements of PMSE at Resolute Bay by Swarnalingam et al. (2011) revealed that, for strong and moderate PMSE, the median aspect angles at 84 km are 2-4° smaller that those at 88 km. Our results show slightly less difference (1-2°) between aspect angles at 82 km and 87 km. This morphology is supported by the results of rocket measurements of turbulence by Lübken et al. (2002). These show that the turbulence occurrence rate starts to grow from 80 km and maximises at about 88 km (Rapp and Lübken, 2003).

Chilson et al. (2002) and Zecha et al. (2001) noticed enhanced aspect sensitivity at the lower edges of PMSE sublayers. Similar features are seen in our Fig. 2. We do not address this topic here because it requires additional investigation, results of which will be reported elsewhere.

Already the first measurements of PMSE aspect sensitivity by Czechowsky et al. (1988) showed its relation to echo power. The authors, basing their findings on the two-day data, reported that the peaks in backscattered power profile correspond to minima in aspect angle profile. Similarly, Chilson et al. (2002) found for one-day data that PMSE regions with enhanced aspect sensitivity (small aspect angles) have high signal-to-noise ratio. However, Swarnalingam et al. (2011) analysed data for 12 years and did not find an obvious correlation between PMSE aspect sensitivity and echo volume reflectivity. This result is in full agreement with our findings.
Swarnalingam et al. mentioned a possible tendency for strong PMSE to have low aspect angles. Similar features can be seen in our Figs. 6 and 7. However, this apparent impression is based on poor statistics available for the strongest PMSE in our paper as well as in Swarnalingam et al. (2011). In some studies (e.g. Hoppe et al., 1990; Blix et al, 1999) there were attempts to relate echo power, spectral width and aspect sensitivity in order to identify turbulent or Fresnel scatter. Although we not analyse PMSE spectral widths in this paper, based on the calculated values of $\theta_s$ we can conclude that at least half of PMSE are more due to Fresnel scattering.

We found that half of all PMSE detected by ESRAD have high aspect sensitivity. If PMSE are strongly aspect sensitive this can affect estimation of their absolute strength expressed in terms of volume reflectivity $\eta$. In the calculation of $\eta$ it is assumed that scatter is isotropic and fills the entire radar sampling volume. However, if the scatterers’ polar diagram is narrower than that of the two-way radar beam then one will underestimate their volume reflectivity. We evaluate this effect for ESRAD using our results on PMSE aspect sensitivity.

Volume reflectivity $\eta$ is defined as total power scattered isotropically by unit scattering volume, per unit transmit power, per unit solid angle. For narrow and moderate beam radars $\eta$ is inversely proportional to the effective radar volume $V_0$, which is approximated as follows (Hocking, 1985):

$$V_0 = \pi h \left(\theta_{TR}\right)^2 \Delta h,$$

where $\theta_{TR}$ is the e\(^{-1}\) half-width of the two-way radar polar diagram, $h$ is altitude and $\Delta h$ is altitude resolution. For aspect sensitive scatter the effective radar volume is determined by $\theta_{\text{eff}}$. We have calculated $\theta_{TR}$ to be 3.6° before the ESRAD antenna extension in 2004, and 2.4° afterward, and median values of $\theta_s$, averaged over the first and the second intervals are 3.2° and 3.1°, respectively. By using Eq. 4 we reached the values of $\theta_{\text{eff}}$ of 2.4° and 1.9° before and after 2004, respectively. Finally, we found that we underestimated volume reflectivity of at least half the PMSE (those which are aspect sensitive) detected with ESRAD by a factor of 1.6-2.3 (2-3.5 dB). By comparison, Swarnalingam et al. (2009b) reported 0.3 dB possible effect due to high aspect sensitivity (5°) evaluated for the Resolute Bay VHF radar with a 1.4° two-way radar beam width.
6. Summary
We used ESRAD PMSE data for 1997-2010 to calculate in-beam PMSE aspect sensitivities for 11 years using the FCA technique. We found that 50 - 52 % of data can be identified as highly aspect sensitive echoes. The rest of PMSE have aspect sensitivity parameters $\theta_s$, characterising the half-width of the scatterers’ polar diagram, larger than 9-11°, and their values cannot be quantified solely using the ESRAD vertical beam measurements.

We calculated the distribution of PMSE over the aspect sensitivity parameter $\theta_s$ for each year and found that they remain consistent from year to year. The median values of $\theta_s$ are in the 2.9-3.7° range. It was found that when calculating volume reflectivity for such aspect sensitive PMSE, one can underestimate it by more than 3 dB.

We found also that $\theta_s$ slightly increases with altitude (by 1-2° from 82 km to 87 km). This altitude dependence of $\theta_s$ does not change from year to year. No dependence of PMSE aspect sensitivity on backscattered power was identified for any year.

We analysed the limitations of the in-beam and off-zenith beam methods related to the finite radar beam width. Our conclusion is that the former is suitable for highly aspect sensitive echoes while the latter is needed for more isotropic scatterers. Both techniques should be combined in order to get full information about PMSE aspect sensitivity.

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References


Kirkwood, S., Belova, E., Satheesan, K., Narayana Rao, T., Rajendra Prasad, T., Satheesh Kumar, S.: Fresnel scatter revisited – comparison of 50 MHz radar and