

## **Authors' response to comments on paper acp-2015-326**

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We thank the referee and the editor for their latest comments. We appreciate the constructive feedback on our manuscript and have made substantial changes. We have focused more on explaining temperature variability in the dataset in the new version and think this has improved the manuscript. We hope the referee and the editor agree on this.

### **Response to anonymous referee # 2:**

1. It is clear that quite a bit of massaging of the data has been necessary to get the temperature time series used in the paper. As far as I can tell, the authors have clearly laid out the steps that they took and the various uncertainties added at each step. I think readers are suitably warned about the reliability of the resulting temperature data and, therefore, I do not have anything more to add about it. Others who analyze similar types of data will be interested in the steps to follow.

*Authors' response:* No response required.

2. It seems that my previous review was not clear enough about my concerns with the solar and trend analysis. It is simply not possible to separate these two sources of change during this period. The short data record would be a serious problem for any analysis that is limited to one solar cycle or less. (Note: despite being 11 years, your data span is substantially shorter than the full cycle between the maxima of SS 23 and 24, hence less than one solar cycle.) Because of the strong difference in the solar flux at the beginning and end of your period, there is a marked trend in the F10.7 flux itself. The atmospheric temperature trend (or change; whichever you call it, it is the same problem) and the response to solar variability cannot be separated. This has nothing to do with the analysis technique you use, it is not possible because the two terms you are using as predictors (solar flux and time) are significantly correlated.

*Authors' response:* The manuscript has now been rewritten substantially and new analyses have been conducted. In the new version, we have investigated temperature variability in the dataset using Lomb-Scargle periodogram analysis and found periodic oscillations which we used for seasonally de-trending the temperature record. Trend has been investigated, but response to solar variability has not been incorporated in the trend analysis. The new analyses have resulted in that the Results and Discussion sections have been rewritten to a large extent.

3. As per the above comment, the paper should not be published with the trend and solar analysis (Sections 4, 5.2, and 6). If you want to show the deseasonalized data along with the F10.7 flux, that would be OK. If so, it is necessary to acknowledge the ambiguity between possible solar cycle and trend variations. For example, time information could be included in

Figure 8 by color-coding based on year. As shown, the figure is not an adequate representation of the solar response and therefore the fits (red and blue lines) are not meaningful. Further analysis to actually define these two responses separately should not be included because it cannot be scientifically supported.

*Authors' response:* Figure 8 (Figure 9 in the new version) is now, instead of showing fits to a scatterplot of NTMR temperature vs F10.7 values, showing the progression of yearly values of temperature residuals and F10.7 from high solar flux values to low values and back to high values. As described above, response to solar variability has not been subtracted from the temperatures before estimating the trend.

Minor comment

1. (p.15292, revision near top) Do you mean decreasing concentrations of O<sub>3</sub>?

*Authors' response:* Changed to “altered concentrations of CO<sub>2</sub> and O<sub>3</sub>”.

### **Additional comments to the referee and editor:**

Numerous corrections have been made to the manuscript text and to the figures. It is difficult to give a detailed overview of all corrections, but we have listed here the ones that have to do with changed content of the text:

1. Some of the figures from the previous version have been omitted in the new version to make room for new ones related to analyses done on temperature variability. Also, the old Figure 4 has been replaced with a table (Table 1). We hope that the total number of figures and tables is satisfactory.
2. We propose a change of title to “Neutral atmosphere temperature trends and variability at 90 km, 70°N, 19°E, 2003-2014”.
3. Standard error of the mean has not in any of the previous versions been plotted as error bars of daily NTMR temperatures (in e.g. Figure 2). This has been done for better legibility of the figures. However, since this is bad practice, we have included information about the standard error of the mean of temperatures in the text instead (page 5, lines 13-15).
4. The length of time series required for detection of a trend has been estimated, following the criterion from Weatherhead et al. (1998) (page 9, lines 5-15; page 10, lines 6-7).

References:

Weatherhead, E. C., G. C. Reinsel, G. C. Tiao, X.-L. Meng, D. Choi, W.-K. Cheang, T. Keller, J. DeLuigi, D. J. Wuebbles, J. B. Kerr, A. J. Miller, S. J. Oltmans, and J. E. Frederick: Factors affecting the detection of trends: Statistical considerations and applications to environmental data, *J. Geophys. Res.* 103 (D14), 17149-17161, doi: 10.1029/98JD00995, 1998.

1 **Neutral atmosphere temperature change at 90 km, 70°N,**  
2 **19°E, 2003-2014**

3

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13


14 **Abstract**

15 Neutral temperatures for 90 km height above Tromsø, Norway, have been determined using  
16 ambipolar diffusion coefficients calculated from meteor echo fading times using the  
17 Nippon/Norway Tromsø Meteor Radar (NTMR). Daily temperature averages have been  
18 calculated from November 2003 to October 2014 and calibrated against temperature  
19 measurements from the Microwave Limb Sounder (MLS) on board Aura. ~~The long-term trend~~  
20 ~~of temperatures from the NTMR radar is investigated, and winter and summer seasons are~~  
21 ~~looked at separately. Seasonal variation has been accounted for, as well as solar response,~~  
22 ~~using the F10.7 cm flux as a proxy for solar activity.~~ The temperature change from 2003 to  
23 2014 is ~~4.1 K ± 1.3 K/decade~~, while in summer and winter the change is ~~0.5 K ± 2.5~~  
24 ~~K/decade~~ and ~~10.6 K ± 3.8 K/decade~~, respectively.  How well suited a meteor radar is for  
25 estimating neutral temperatures at 90 km using meteor trail echoes is discussed, and physical  
26 explanations behind a cooling trend are proposed.

# 1 1 Introduction

2 Temperature changes in the mesosphere and lower thermosphere (MLT) region due to both  
3 natural and anthropogenic variations cannot be assessed without understanding the dynamical,  
4 radiative and chemical couplings between the different atmospheric layers. Processes  
5 responsible for heating and cooling in the MLT region are many. Absorption of UV by O<sub>3</sub> and  
6 O<sub>2</sub> causes heating, while CO<sub>2</sub> causes strong radiative cooling. Planetary waves (PWs) and  
7 gravity waves (GWs) break and deposit heat and momentum into the middle atmosphere and  
8 influence the mesospheric residual circulation, which is the summer-to-winter circulation in  
9 the mesosphere. Also, heat is transported through advection and adiabatic processes.

10 For decades, it has been generally accepted that increased anthropogenic emissions of  
11 greenhouse gases are responsible for warming of the lower atmosphere (e.g. Manabe and  
12 Wetherald, 1975), and that these emissions are causing the mesosphere and thermosphere to  
13 cool (Akmaev and Fomichev, 2000; Roble and Dickinson, 1989). Akmaev and Fomichev  
14 (1998) reported, using a middle atmospheric model, that if CO<sub>2</sub> concentrations are doubled,  
15 temperatures will decrease by about 14 K at the stratopause, by about 10 K in the upper  
16 mesosphere and by 40-50 K in the thermosphere. Newer and more sophisticated models  
17 include important radiative and dynamical processes as well as interactive chemistries. Some  
18 model results indicate a cooling rate near the mesopause less than predicted by Akmaev and  
19 Fomichev (1998), while others maintain the negative signal (French and Klekociuk, 2011;  
20 Beig, 2011a). The thermal response in this region is strongly influenced by changes in  
21 dynamics, and some dynamical processes contribute to a warming which counteracts the  
22 cooling expected from greenhouse gas emissions (Schmidt et al., 2006).

23 Even though the increasing concentration of greenhouse gases is generally accepted to be the  
24 main driver, also other drivers of long-term changes and temperature trends exist, namely  
25 stratospheric ozone depletion, long-term changes of solar and geomagnetic activity, secular  
26 changes of the Earth's magnetic field, long-term changes of atmospheric circulation and  
27 mesospheric water vapour concentration (Laštovička et al., 2012)  the complexity of  
28 temperature trends in the MLT region and their causes act as motivation for studying these  
29 matters further.

30 In this paper, we investigate long-term change of temperatures obtained from the NTMR  
31 radar, and we also look at summer and winter seasons separately. In Sect. 2, specifications of  
32 the NTMR radar are given, and the theory behind the retrieval of temperatures using

1 ambipolar diffusion coefficients from meteor trail echoes is explained. In Sect. 3, the method  
2 behind the calibration of NTMR temperatures against Aura MLS temperatures is explained.  
3 Section 4 treats ~~the temperature trend analysis, including the correction for seasonal variation~~  
4 ~~and solar response~~. The theory and underlying assumptions for the method of determining  
5 neutral temperatures from meteor trail echoes and thus how well suited a meteor radar is for  
6 estimating such temperatures is discussed in Sect. 5. Also, physical explanations behind the  
7 change in temperature are discussed, as well as comparison with other reports on trends.

## 8 **2 Instrumentation and data**

9 The Nippon/Tromsø Meteor Radar (NTMR) is located at Ramfjordmoen near Tromsø, at  
10 69.58°N, 19.22°E. It is operated 24 hours a day, all year round. Measurements are available  
11 for more than 90 % of all days since the radar was first operative in November 2003. The  
12 meteor radar consists of one transmitter antenna and five receivers and is operating at 30.25  
13 MHz. It detects echoes from ionized trails from meteors, which appear when meteors enter  
14 and interact with the Earth's neutral atmosphere in the MLT region. The ionized atoms from  
15 the meteors are thermalized, and the resulting trails expand in the radial direction mainly due  
16 to ambipolar diffusion, which is diffusion in plasma due to interaction with the electric field.  
17 Underdense meteors, which are the ones used in this study, have a plasma frequency that is  
18 lower than the frequency of the radar, which makes it possible for the radio wave from the  
19 radar to penetrate into the meteor trail and be scattered by each electron.

20 Echoes are detected from a region within a radius of approximately 100 km (horizontal  
21 space). The radar typically detects around 10000 echoes a day, of which around 200-600  
22 echoes are detected per hour at the peak occurrence height of 90 km. Figure 1 shows the  
23 vertical distribution of meteor echoes as a function of height, averaged over the time period  
24 2003-2014. The number of echoes detected per day allows for a 30 minute resolution of  
25 temperature values. The intra-day periodicity in meteor detections by the NTMR radar is less  
26 pronounced than that of lower latitude stations and we do not anticipate tidally-induced bias  
27 regarding echo rates at specific tidal phases for daily averages. The height resolution and the  
28 range resolution are both 1 km, when looking at altitudes around the peak occurrence height.  
29 From the decay time of the radar signal we can derive ambipolar diffusion coefficients,  $D_a$ :

$$30 \quad D_a = \frac{\lambda^2}{16\pi^2\tau} \quad (1)$$

1 where  $\lambda$  is the radar wavelength and  $\tau$  is the radar echo decay time. It has been shown that this  
2 coefficient also can be expressed in terms of atmospheric temperature and pressure:

$$3 \quad D_a = 6.39 \times 10^{-2} K_0 \frac{T^2}{p} \quad (2)$$

4 where  $p$  is pressure,  $T$  is temperature, and  $K_0$  is the zero-field reduced mobility factor of the  
5 ions in the trail. In this study we use the value for  $K_0$  of  $2.4 \times 10^{-4} \text{ m}^2 \text{ s}^{-1} \text{ V}^{-1}$ , in accordance  
6 with e.g. Holdsworth et al. (2006). Pressure values are derived from atmospheric densities  
7 obtained from falling sphere measurements appropriate for  $70^\circ\text{N}$ , combining those of Lübken  
8 and von Zahn (1991) and Lübken (1999), previously used by e.g. Holdsworth (2006) and  
9 Dyrland et al. (2010). These densities do not take into account long-term solar cycle  
10 variations.


11 The NTMR radar is essentially identical to the Nippon/Norway Svalbard Meteor Radar  
12 (NSMR) located in Adventdalen on Spitsbergen at  $78.33^\circ\text{N}$ ,  $16.00^\circ\text{E}$ . Further explanation of  
13 the radar and explanation of theories can be found in e.g. Hall et al. (2002; 2012), Cervera and  
14 Reid (2000) and McKinley (1961).

15 Calibration of temperatures derived from meteor echoes with an independent, coinciding  
16 temperature series is necessary, according to previous studies (e.g. Hocking, 1999).  
17 Temperatures from the NSMR radar have been derived most recently by Dyrland et al.  
18 (2010), employing a new calibration approach for the meteor radar temperatures, wherein  
19 temperature measurements from the Microwave Limb Sounder (MLS) on the Aura satellite  
20 were used instead of the previously used rotational hydroxyl and potassium lidar temperatures  
21 from ground-based optical instruments (Hall et al., 2006). Neither ground-based optical  
22 observations nor lidar soundings are available for the time period of interest or the location of  
23 the NTMR. In this study we therefore employ the same approach as Dyrland et al. (2010),  
24 using Aura MLS temperatures to calibrate the NTMR temperatures.

25 NASA's EOS Aura satellite was launched 15 July 2004 and gives daily global coverage  
26 (between  $82^\circ\text{S}$  and  $82^\circ\text{N}$ ) with about 14.5 orbits per day. The MLS instrument is one of four  
27 instruments on Aura and samples viewing forward along the spacecraft's flight direction,  
28 scanning its view from the ground to  $\sim 90$  km every  $\sim 25$  seconds, making measurements of  
29 atmospheric temperature, among others (NASA Jet Propulsion Laboratory).

1 Because of a general cooling of most of the stratosphere and mesosphere the last decades due  
2 to e.g. ~~increasing~~ concentrations of CO<sub>2</sub> and O<sub>3</sub>, the atmosphere has been shrinking, leading to  
3 a lowering of pressure surfaces at various altitudes. It is important to distinguish between  
4 trends on fixed pressure altitudes and fixed geometric altitudes, since trends on geometric  
5 altitudes include the effect of a shrinking atmosphere (Lübken et al., 2013). In this study, we  
6 have obtained Aura MLS temperature data (version 3.3) for latitude 69.7°N ± 5.0° and  
7 longitude 19.0°E ± 10.0° at 90 km geometric altitude.

### 8 **3 Calibration of NTMR temperatures**

9 Figure 2 shows daily NTMR temperatures from November 2003 to October 2014, derived  
10 from Eqs. (1) and (2), plotted together with Aura MLS temperature  The Aura satellite  
11 overpasses Tromsø at 01-03 UTC and 10-12 UTC, which means that the Aura daily averages  
12 are representative for these time windows. It was therefore necessary to investigate any bias  
13 arising from Aura not measuring throughout the whole day. A way to do this is to assume that  
14 Aura temperatures and NTMR temperatures follow the same diurnal variation and thus  
15 investigate the diurnal variation of NTMR temperatures. This was done by superposing all  
16 NTMR temperatures by time of day, obtaining 48 values for each day, since the radar allows  
17 for a 30-minute resolution.

18 There is an ongoing investigation into the possibility that D<sub>a</sub> derived by NTMR can be  
19 affected by modified electron mobility during auroral particle precipitation. According to  
20 Rees et al. (1972), neutral temperatures in the auroral zone show a positive correlation with  
21 geomagnetic activity. It is therefore a possibility that apparent D<sub>a</sub> enhancements during strong  
22 auroral events do not necessarily depict neutral temperature increase. This matter requires  
23 further attention.

24 Investigation of possible unrealistic D<sub>a</sub> enhancements was carried out by calculating standard  
25 errors of estimated half hourly D<sub>a</sub> values:

$$26 \quad se = \frac{\sigma}{\sqrt{ne}} \quad (3)$$

27 where  $\sigma$  is standard deviation and  $ne$  is the number of echoes detected by the radar. By  
28 inspection and comparison of results between one of the authors (MT) and Satonori Nozawa  
29 (private comm.), all half hourly D<sub>a</sub> values with a standard error larger than 7 % of the  
30 estimated D<sub>a</sub> value were excluded from further analysis. This rejection criterion led to that 5.4

1 % of the  $D_a$  values were rejected. NTMR temperatures after application of the  $D_a$  rejection  
2 procedure will henceforward be referred to as  $D_a$ -rejected NTMR temperatures.

3 Figure 3 shows monthly averages of the superposed values of  $D_a$ -rejected NTMR  
4 temperatures as a function of time of day for days coinciding with Aura measurements. It is  
5 evident from the figure that the lowest temperatures are in general achieved in the forenoon,  
6 which coincides with one of the periods per day when Aura MLS makes measurements over  
7 Tromsø.

8 Subtracting the monthly averages of the 00-24 UTC temperatures from the 01-03 UTC and  
9 10-12 UTC temperatures gave the estimated biases in Aura daily means due to only sampling  
10 during some hours of the day and are given in Fig. 4. ~~The figure shows that~~ by judging by the  
11 measurement windows, Aura underestimates the daily mean (00-24 UTC) more during winter  
12 than during spring and summer. Note the higher standard deviations in spring and summer  
13 compared to winter.

14 The initially obtained Aura temperatures were corrected by adding the biases from Fig. 3 in  
15 order to arrive at daily mean temperatures that are representative for the entire day. Also, the  
16 Aura temperatures were corrected for “cold bias”. French and Mulligan (2010) reported that  
17 Aura MLS temperatures exhibit a 10 K cold bias compared with OH\*(6-2) nightly  
18 temperatures at Davis Station, Antarctica. A newer study by Garcia-Comas et al. (2014)  
19 shows that Aura MLS exhibits a bias compared with several satellite instruments which varies  
20 with season. According to their findings, a 10 K correction for cold bias was applied to the  
21 Aura summer and winter temperatures (Jun – Aug, Dec - Feb), while a 5 K correction was  
22 applied to autumn and spring temperatures (Sep – Nov, Mar – May). The corrected Aura  
23 temperatures will henceforward be referred to as local time and cold bias-corrected Aura MLS  
24 temperatures.

~~25 Figure 5 shows a scatterplot of the local time and cold bias corrected Aura temperatures~~  
~~26 against the  $D_a$  rejected NTMR temperatures. The red line represents the linear fit ( $R^2 = 0.83$ )~~  
~~27 and is described by:~~

$$28 \quad T_{NTMR} = 0.84T_{Aura} + 32 \quad (4)$$

29 where  $T_{NTMR}$  is the  $D_a$ -rejected temperature obtained from NTMR, and  $T_{Aura}$  is the local time  
30 and cold bias-corrected temperature from Aura MLS. Inverting Eq. (4) enabled us to estimate  
31 NTMR temperatures calibrated with respect to Aura MLS temperatures. NTMR temperatures



1 were now corrected for the days of measurements coinciding with Aura measurements and are  
2 henceforward referred to as MLS-calibrated NTMR temperatures. For calibration of NTMR  
3 temperatures from November 2003 to August 2004 (before the start of the Aura MLS  
4 dataset), the same equation (Eq. 4) was used, using NTMR Da-rejected temperatures from  
5 November 2003 to August 2004 as input instead of  $T_{Aura}$ .

6 To estimate the calibration uncertainty, all local time and cold bias-corrected Aura  
7 temperatures were subtracted from the MLS-calibrated NTMR temperatures, and the  
8 differences were plotted in a histogram with 5 K bins. A Gaussian was fitted to the  
9 distribution. The standard deviation of the Gaussian was 11.9 K, which is then considered the  
10 overall uncertainty of the calibration. Figure 6 shows the histogram and the fitted Gaussian  
11 curve. Finally, Fig. 7 shows the MLS-calibrated NTMR temperatures with uncertainties  
12 plotted together with Aura MLS temperatures, corrected for cold and time-of-day  
13 measurement bias.

#### 14 **4 Trend analysis**

15 ~~A monthly climatology of the MLS-calibrated NTMR temperatures was obtained by~~  
16 ~~averaging all January, February, etc. values. The seasonal variation is not shown here but~~  
17 ~~reveals a summer minimum of around 150-160 K and a winter maximum of around 200-210~~  
18 ~~K. The monthly values were then subtracted from the daily MLS-calibrated temperatures,~~  
19 ~~obtaining daily residuals independent of seasonal variation.~~

20 ~~There are several measures of solar variability available, e.g. the F10.7 cm solar radio flux,~~  
21 ~~the sunspot number (SSN), total solar irradiance (TSI), Mg II 280 nm core-to-wing ratio UV~~  
22 ~~index and the flare index (FI). These indices are considered proxies for solar radiation formed~~  
23 ~~on different altitudes of the solar atmosphere and are highly correlated (Bruevich et al., 2014).~~  
24 ~~In this study we use the F10.7 cm flux as a proxy for solar activity, which is the most~~  
25 ~~commonly used index in middle/upper atmospheric temperature trend studies (e.g. Laštovička~~  
26 ~~et al., 2008; Hall et al., 2012).~~

27 ~~A 30 day running mean filter was applied to the daily residual temperatures. Figure 8 shows~~  
28 ~~the residuals plotted against corresponding F10.7 cm values. The straight, red line in the~~  
29 ~~figure gave the best linear fit to the daily residuals with a 30 day running mean applied and~~  
30 ~~gave a solar response coefficient of  $4.5 \text{ K} \pm 0.3 \text{ K}/100 \text{ SFU}$  (1 SFU = 1 solar flux unit =  $10^{-22}$~~   
31  ~~$\text{W m}^{-2} \text{ Hz}^{-1}$ ).~~

1 From Fig. 8 there appears to be a somewhat non-linear relationship between the temperatures  
2 at 90 km height and the F10.7 cm index. There seems to be a tendency of a less steep increase  
3 in temperatures toward higher F10.7 values. Ogawa et al. (2014) also found a non-linear  
4 relationship between upper atmospheric temperatures and solar activity using EISCAT UHF  
5 radar observations from 200 to 450 km altitude over Tromsø, even though it must be noted  
6 that the altitude range they looked at differs from ours. In Fig. 8 we have therefore also  
7 plotted the quadratic, least-squares fit to the running mean values. Lacking any objective  
8 scientific basis to do otherwise, we chose to fit a 2<sup>nd</sup> degree polynomial following the  
9 philosophy of Ogawa et al. (2014), although it is conceivable that other functions could be  
10 more suitable. The 2<sup>nd</sup> degree polynomial gave us a better fit to the residuals ( $R^2=0.16$ )  
11 compared to the straight line ( $R^2=0.03$ ). We subtracted the solar response from the dataset of  
12 daily, seasonally corrected residuals using this relation:

$$13 \quad T' = T - (a \cdot f10.7^2 + b \cdot f10.7 + c) \quad (5)$$

14 where  $T'$  is the new set of residual temperatures with seasonal and solar response subtracted,  
15  $T$  is the residual temperatures with only seasonal variation subtracted,  $f10.7$  is the daily F10.7  
16 cm flux corresponding to  $T$ , and  $a$ ,  $b$  and  $c$  are coefficients of the 2<sup>nd</sup> degree polynomial ( $a=$   
17  $0.0017$ ,  $b=0.41$ ,  $c=23$ ).

18 From the new set of temperature residuals we calculated monthly means. This was done to  
19 remove any high frequency deterministic component, such as that resulting from multi day  
20 period waves. Finally, the linear trend was found by performing linear regression using a  
21 least-squares fit. The long-term linear temperature trend using monthly means is 4.1 K with a  
22 standard error of 1.3 K/decade. This trend can be considered statistically significant (i.e.  
23 significantly non-zero at the 5 % level), since the uncertainty ( $2\sigma = 2.6$  K/decade) is less than  
24 the trend itself (Tiao et al., 1990). Figure 9 shows the linear trend of the monthly values for  
25 the whole dataset, from November 2003 through October 2014. For comparison, the trend  
26 when applying the linear solar response using monthly values is  $4.7 \text{ K} \pm 1.4 \text{ K/decade}$ , and  
27 the trend using daily temperature values (applied Eq. 5 for subtraction of solar response) is  
28  $3.8 \text{ K} \pm 0.6 \text{ K/decade}$ .

29 In addition to the average temperature change, we also treated summer and winter seasons  
30 separately. First, trends for each month were investigated using the same approach as for the  
31 average regardless of month. Figure 10 shows the result. Then, averages of November,  
32 December and January, and of May, June and July were made. They were defined as “winter”

1 and “summer”, respectively. The long-term linear winter trend is  $10.6 \text{ K} \pm 3.8 \text{ K/decade}$ , and  
2 the long-term summer trend is  $-0.5 \text{ K} \pm 2.5 \text{ K/decade}$ .

3 The trend analysis was also performed without carrying out the  $D_a$  rejection procedure  
4 explained in Sect. 3. Final results with and without data rejection do not differ significantly  
5 considering the calculated uncertainties.

## 6 5 Discussion



### 7 5.1 Suitability of a meteor radar for estimation of neutral temperatures at 90 8 km height

9 As explained in Sect. 2, neutral air temperatures derived from meteor trail echoes depend on  
10 pressure,  $p$ , the zero-field reduced mobility of the ions in the trail,  $K_0$ , and ambipolar diffusion  
11 coefficients,  $D_a$ .  $K_0$  will depend on the ion composition in the meteor trail, as well as the  
12 chemical composition of the atmosphere. The chemical composition of the atmosphere is  
13 assumed to not change significantly with season (Hocking, 2004). Unfortunately, the exact  
14 content of a meteor trail is unknown. Usually, a value for  $K_0$  between  $1.9 \cdot 10^{-4} \text{ m}^2 \text{ s}^{-1} \text{ V}^{-1}$  and  
15  $2.9 \cdot 10^{-4} \text{ m}^2 \text{ s}^{-1} \text{ V}^{-1}$  is chosen, depending on what ion one assumes to be the main ion of the  
16 trail (Hocking et al., 1997). Even though we in this study have chosen a constant value for  $K_0$   
17 of  $2.4 \cdot 10^{-4} \text{ m}^2 \text{ s}^{-1} \text{ V}^{-1}$ , some variability in  $K_0$  is expected. According to Hocking (2004)  
18 variability can occur due to fragmentation of the incoming meteoroid, anisotropy in the  
19 diffusion rate, plasma instabilities and variations in the composition of the meteor trail. Using  
20 computer simulations, they reported a typical variability in  $K_0$  from meteor to meteor of 27 %  
21 and that the variability is most dominant at higher temperatures. Based on this, we cannot rule  
22 out sources of error due to the choice of  $K_0$  as a constant, but since we have no possibility to  
23 analyse the composition of all meteor trails detected by the radar we have no other choice  
24 than to choose a constant value for  $K_0$ .

25 How well ambipolar diffusion coefficients obtained for 90 km altitude are suited for  
26 calculating neutral temperatures has previously been widely discussed, e.g. by Hall et al.  
27 (2012) for the trend analysis of the Svalbard meteor radar data, but will be shortly repeated  
28 here. For calculations of temperatures using meteor radar, ambipolar diffusion alone is  
29 assumed to determine the decay of the underdense echoes. Diffusivities are expected to  
30 increase exponentially with height through the region from which meteor echoes are obtained  
31 (Ballinger et al., 2008; Chilson et al., 1996). Hall et al. (2005) found that this is only the case

1 between ~85 and ~95 km altitude, using diffusion coefficients delivered by NTMR from  
2 2004. They found diffusivities less than expected above ~95 km and diffusivities higher than  
3 expected below ~85 km. Ballinger et al. (2008) got a similar result using meteor observations  
4 over northern Sweden. It has been proposed that processes other than ambipolar diffusion  
5 influence meteor decay times. If this is the case it may have consequences for the estimation  
6 of temperatures, and therefore it is important to investigate this further.

7 Departures of the anticipated exponential increase with height of molecular diffusion above  
8 ~95 km have in previous studies been attributed to gradient-drift Farley-Buneman instability.  
9 Farley-Buneman instability occurs where the trail density gradient and electric field are  
10 largest. Due to frequent collisions with neutral particles, electrons are magnetised while ions  
11 are left unmagnetised, causing electrons and ions to differ in velocity. Electrons then create an  
12 electric field perpendicular to the meteor trail, leading to anomalous fading times that can be  
13 an order of magnitude higher than those expected from ambipolar diffusion. The minimum  
14 altitude at which this occurs depends on the trail altitude, density gradient and latitude, and at  
15 high latitudes this altitude is ~95 km. Therefore, using ambipolar diffusion rates to calculate  
16 trail altitudes above this minimum altitude may lead to errors of several kilometres, due to  
17 that the diffusion coefficients derived from the measurements are underestimated (Ballinger et  
18 al., 2008; Dyrud et al., 2001; Kovalev et al., 2008).

19 Reasons for the higher diffusivities than expected according to theory below ~85 km are not  
20 completely understood. Hall (2002) proposed that neutral turbulence may be responsible for  
21 overestimates of molecular diffusivity in the region ~70-85 km, but this hypothesis was  
22 rejected by Hall et al. (2005) due to a lacking correlation between neutral air turbulent  
23 intensity and diffusion coefficients delivered by the NTMR radar. Other mechanisms for  
24 overestimates of molecular diffusivity include incorrect determination of echo altitude and  
25 fading times due to limitations of the radar (Hall et al., 2005).

26 Since the peak echo occurrence height is 90 km and this is also the height at which a  
27 minimum of disturbing effects occur, 90 km height is therefore considered the optimal height  
28 for temperature measurements using meteor radar. Ballinger et al. (2008) report that meteor  
29 radars in general deliver reliable daily temperature estimates near the mesopause using the  
30 method outlined in this study, but emphasize that one should exercise caution when assuming  
31 that observed meteor echo fading times are primarily governed by ambipolar diffusion. They  
32 proposed, after Havnes and Sigernes (2005), that electron-ion recombination can impact


1 meteor echo decay times. Especially can this affect the weaker echoes, and hence can this  
2 effect lead to underestimation of temperatures.


3 Determination of temperatures from meteor radar echo times is a non-trivial task, mainly  
4 because the calculation of ambipolar diffusion coefficients depends on the ambient  
5 atmospheric pressure. By using radar echo decay times to calculate ambipolar diffusion  
6 coefficients from Eq. 1, we can from Eq. 2 get an estimate for  $T^2/p$ . Input of pressure values  
7 into the equation will thus provide atmospheric temperatures. However, measurements of  
8 pressure are rare and difficult to achieve at 90 km height, and often one has to rely on model  
9 values. Traditionally, pressure values at 90 km have been calculated using the ideal gas law,  
10 taking total mass density from atmospheric models, e.g. the MSISE models, where the newest  
11 version is NRLMSIS-00. It is hard to verify the pressure values derived from the models  
12 because of lack of measurements to compare the model to, and hence using the pressure  
13 values may result in uncertainties of estimated atmospheric temperatures. In this study, we  
14 obtained pressure values from measurements of mass densities obtained from falling spheres  
15 combined with sodium lidar from Andøya (69°N, 15.5°E) (Lübken, 1999; Lübken and von  
16 Zahn, 1991). All measurements have been combined to give a yearly climatology, that is, one  
17 pressure value for each day of the year. Since Andøya is located in close proximity to Tromsø  
18 (approximately 120 km), the pressure values are considered appropriate for our calculations of  
19 neutral temperatures. One disadvantage with using pressure values obtained from the falling  
20 sphere measurements is that no day-to-day variations are taken into account, only the average  
21 climatology.

## 22 **5.2 Physical explanations for cooling and comparison with other studies**

23 Other studies on long-term mesospheric temperature trends from mid and high latitudes yield  
24 mostly negative or near-zero trends. Few studies cover the same time period as ours, and few  
25 are from locations close to Tromsø. Hall et al. (2012) reported a negative trend of  $-4 \text{ K} \pm 2$   
26  $\text{K/decade}$  for temperatures derived from the meteor radar from Longyearbyen, Svalbard  
27 (78°N, 16°E) at 90 km height over the time period 2001 to 2011, while Holmen et al. (2014)  
28 found a near-zero trend for OH\* airglow temperatures at ~87 km height over Longyearbyen  
29 over the longer time period 1983 to 2013. Offermann et al. (2010) reported a trend of  $-2.3 \text{ K} \pm$   
30  $0.6 \text{ K/decade}$  for ~87 km height using OH\* airglow measurements from Wuppertal (51°N,  
31 7°E). It must be noted that the peak altitude of the OH\* airglow layer can vary and thus affect  
32 the comparability of OH\* airglow temperature trends and meteor radar temperature trends.

1 Winick et al., 2009 report that the OH\* airglow layer can range from 75 to >90 km, while the  
2 newer study by von Savigny, 2015, indicates that the layer height at high latitudes is  
3 remarkably constant from 2003 to 2011. Beig (2011a) reported that most recent studies on  
4 mesopause region temperature trends show weak negative trends, which is in line with our  
5 results.

6  ~~The estimated solar response coefficient of  $4.5 \text{ K} \pm 0.3 \text{ K}/100 \text{ SFU}$  is somewhat smaller than  
7 values obtained by Forbes et al. (2014) using SABER data from 2002 through 2013. For  
8 latitudes  $50^\circ\text{N}$ – $80^\circ\text{N}$  and altitudes 90–100 km solar response coefficients obtained by Forbes  
9 et al. are close to  $10 \text{ K}/100 \text{ SFU}$ . However, Beig (2011b) summarizes results from numerous  
10 studies on temperature variability due to solar activity and reports that temperature response  
11 in general is  $4\text{--}5 \text{ K}/100 \text{ SFU}$  in the upper part of the mesopause. Our result is in line with this.~~

12 Our results indicate a cooling at 90 km altitude over Tromsø  A general cooling of the middle  
13 atmosphere will cause a contraction of the atmospheric column and hence a lowering of upper  
14 mesospheric pressure surfaces. The pressure model used as input to Eq. 2 is only seasonally  
15 dependent, so a possible trend in pressure at 90 km must be addressed. By looking at Eq. 2, it  
16 is evident that if pressure decreases, temperature will decrease even more. By incorporating a  
17 decreasing trend in the pressure model will then serve to further strengthen the negative  
18 temperature trend we observe.

19 It has been proposed that GWs may be a major cause of negative temperature trends in the  
20 mesosphere and thermosphere (Beig, 2011a; Oliver et al., 2013). GWs effectively transport  
21 chemical species and heat in the region, and increased GW drag leads to a net effect of  
22 cooling above the turbopause (Yigit and Medvedev, 2009). GWs are shown to heat the  
23 atmosphere below about 110 km altitude, while they cool the atmosphere at higher altitudes  
24 by inducing a downward heat flux (Walterscheid, 1981). However, there are large regional  
25 differences regarding trends in GW activity. Hoffmann et al. (2011) found an increasing GW  
26 activity in the mesosphere in summer for selected locations, but Jacobi (2014) found larger  
27 GW amplitudes during solar maximum and related this to a stronger mesospheric jet during  
28 solar maximum, both for winter and summer. Since we have not conducted any gravity wave  
29 trend assessment in this study, we cannot conclude that GW activity is responsible for the  
30 negative temperature trend, but we cannot rule out its role either. ~~It is not the purpose of this  
31 study to analyse the contribution of dynamics to the temperature change we observe.~~

1 The stronger cooling trend for winter compared to summer is consistent with model studies.  
2 Schmidt et al. (2006) and Fomichev et al. (2007) show, using the HAMMONIA and CMAM  
3 models, respectively, that a doubling of the CO<sub>2</sub> concentration will lead to a general cooling  
4 of the middle atmosphere, but that the high-latitude summer mesopause will experience  
5 insignificant change or even slight warming. They propose that this is the result of both  
6 radiative and dynamical effects. In summer, the CO<sub>2</sub> radiative forcing is positive due to heat  
7 exchange between the cold polar mesopause and the warmer, underlying layers. Also, CO<sub>2</sub>  
8 doubling alters the mesospheric residual circulation. This change is caused by a warming in  
9 the tropical troposphere and cooling in the extratropical tropopause, leading to a stronger  
10 equator-to-pole temperature gradient and hence stronger midlatitude tropospheric westerlies.  
11 This causes the westerly gravity wave drag to weaken, resulting in decreased adiabatic  
12 cooling from a slower ascent of the upper mesospheric circulation.

~~13 Our results indicate a cooling at 90 km altitude over Tromsø between 2003 and 2014 when  
14 deseasonalizing our dataset and calculating trend using all months of the year. At the same  
15 time we get quite different trends for winter and summer. We emphasize that this may give a  
16 result that may be somewhat difficult to interpret.~~

17

## 18 **6 Conclusions**



~~19 The long term trend of neutral temperatures at 90 km height derived from the NTMR radar in  
20 Ramfjordmoen, Tromsø, with seasonality and solar response subtracted, is  $-4.1 \text{ K} \pm 1.3$   
21  $\text{K/decade}$ . The linear fit between the smoothed daily residuals and corresponding F10.7 cm  
22 values gave a solar response coefficient of  $4.5 \text{ K} \pm 0.3 \text{ K/100 SFU}$ . However, a 2<sup>nd</sup> degree  
23 polynomial gave the best fit to the data and was thus used for correcting the dataset of solar  
24 response. When looking at summer and winter seasons separately, the trends are  $-0.5 \text{ K} \pm 2.5$   
25  $\text{K/decade}$  for summer and  $-10.6 \text{ K} \pm 3.8 \text{ K/decade}$  for winter.~~

~~26 Final results of the trend analysis, both when excluding and including rejection of  $D_a$  values  
27 due to hypothetical anomalous electrodynamic processes, do not differ significantly. It is  
28 reasonable to believe that strong geomagnetic conditions can affect derived temperatures on a  
29 short time scale. However, due to the considerable quantity of data employed in this study, it  
30 is inconceivable that this effect will change the conclusions regarding trends, as our results  
31 also show.~~

~~1 90 km is considered the optimal height for retrieval of neutral temperatures using ambipolar  
2 diffusion coefficients from NTMR, due to that the peak echo occurrence height detected by  
3 the radar is 90 km and that this is also the height at which a minimum of disturbing effects  
4 occur. Above 95 km anomalous fading times that can be an order of magnitude higher than  
5 those expected from ambipolar diffusion may be measured, due to gradient drift Farley-  
6 Buneman instability, causing the derived ambipolar diffusion to be underestimated. Below  
7 85 km higher diffusivities than expected according to theory in which the temperature  
8 estimation is based on may be encountered, due to reasons not fully understood.~~

9 A weak cooling trend is in line with other recent studies on mesopause region temperature  
10 trends. A cooling of the middle atmosphere will cause a lowering of upper mesospheric  
11 pressure surfaces. By implementing a negative trend in pressure at 90 km into the equation we  
12 use for estimating temperatures the negative temperature trend is enhanced, which reinforces  
13 our finding of a cooling trend. The most accepted theory behind a cooling of the middle  
14 atmosphere is increased greenhouse gas emissions, but also dynamics may play a significant  
15 role. Our results yield a more negative trend in winter compared to summer, which may be  
16 explained by both radiative and dynamical effects. In summer, a larger heat exchange takes  
17 place from atmospheric layers below the cold, polar mesopause. Weakening of gravity wave  
18 drag leads to weakening of the mesospheric residual circulation, which counteracts cooling.  
19 These effects occur due to increased CO<sub>2</sub> concentrations in the atmosphere, according to  
20 model studies.

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25 Maynooth University, Ireland, for providing the NASA EOS Aura MLS temperatures.



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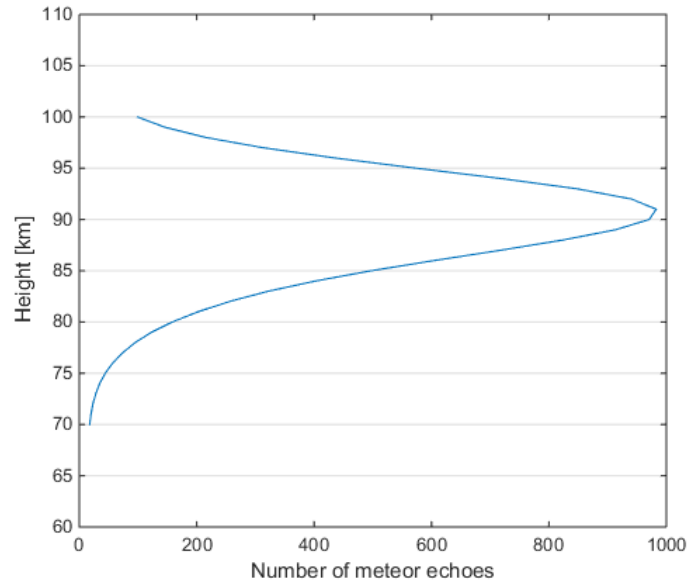
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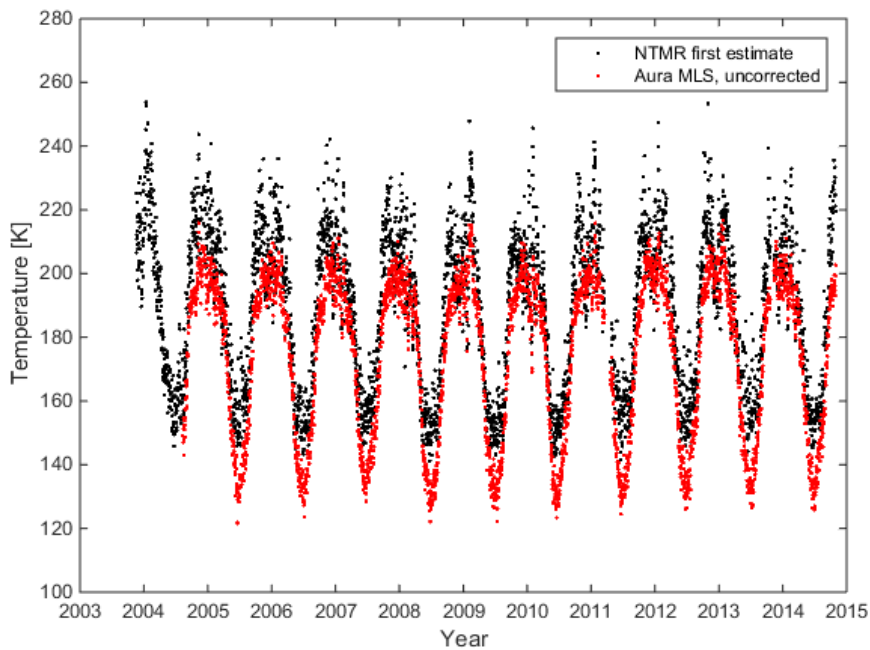


1

2 Figure 1: Vertical distribution of the occurrence of meteor echoes over Tromsø, averaged over  
 3 height between 2003 and 2014. The peak occurrence height is just over 90 km altitude.

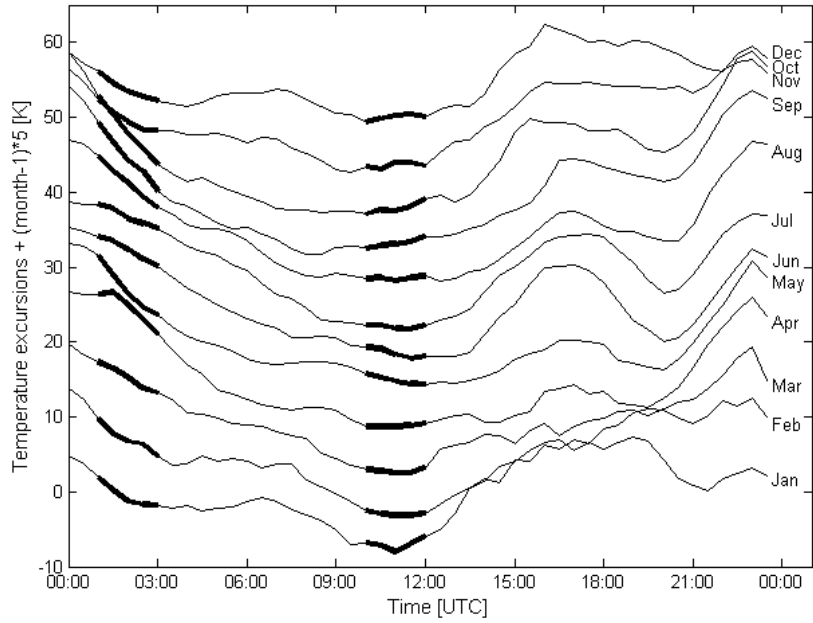
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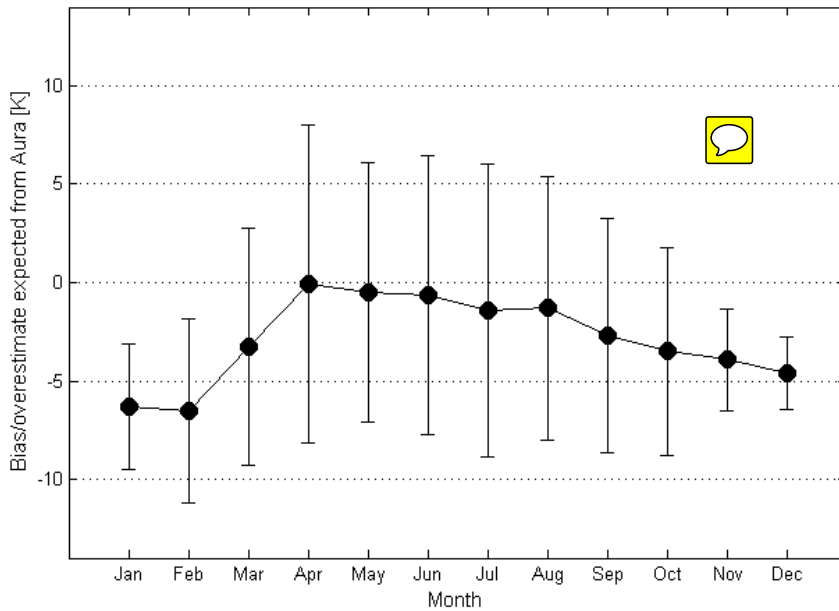


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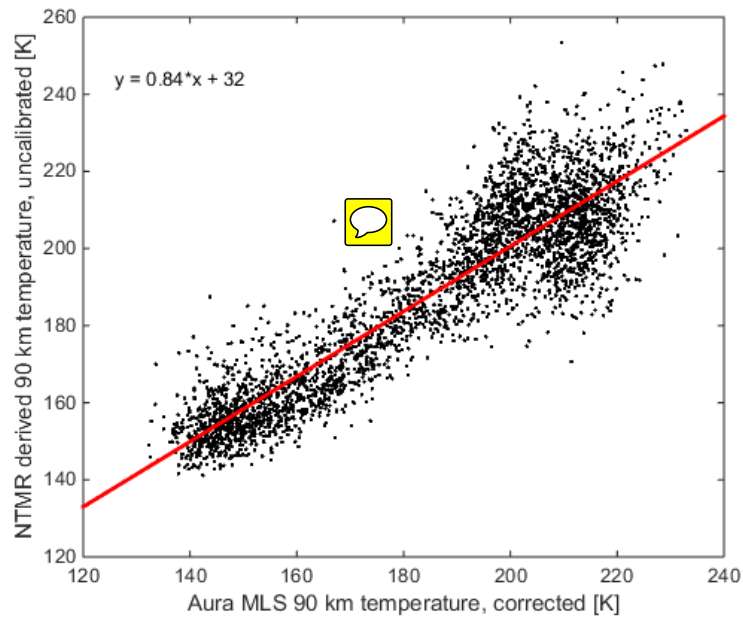
7 Figure 2. Daily values of NTMR temperatures derived from Eqs. (1) and (2), before  
 8 correction for high  $D_a$ , plotted together with Aura MLS temperatures, before applying any  
 9 corrections.



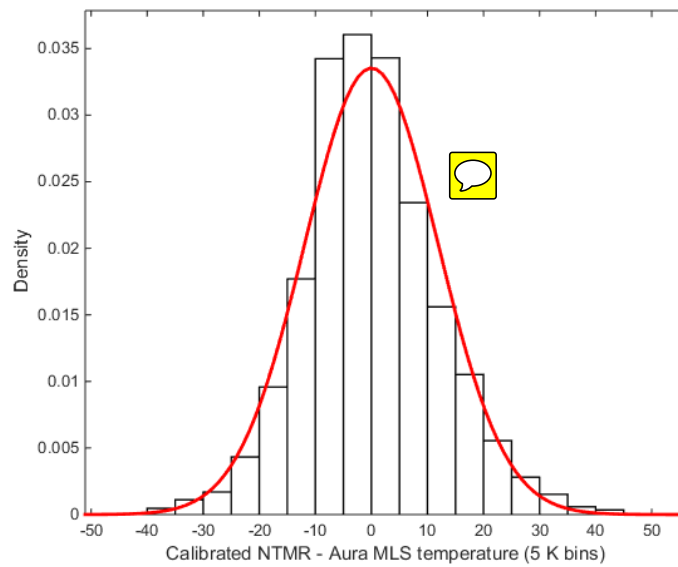
1  
 2 Figure 3. Monthly averages of diurnal temperature variation derived from NTMR after  
 3 correction for high  $D_a$  at 90 km altitude. For clarity time series are displaced by 5 K per  
 4 month subsequent to January. The time of day corresponding to when Aura makes  
 5 measurements over Tromsø (01-03 UTC and 10-12 UTC) is highlighted.



6  
 7 ~~Figure 4. Bias in Aura monthly averages due to that Aura MLS only measures between 01~~  
 8 ~~UTC and 03 UTC, and between 10 UTC and 12 UTC. Error bars represent standard~~  
 9 ~~deviations.~~

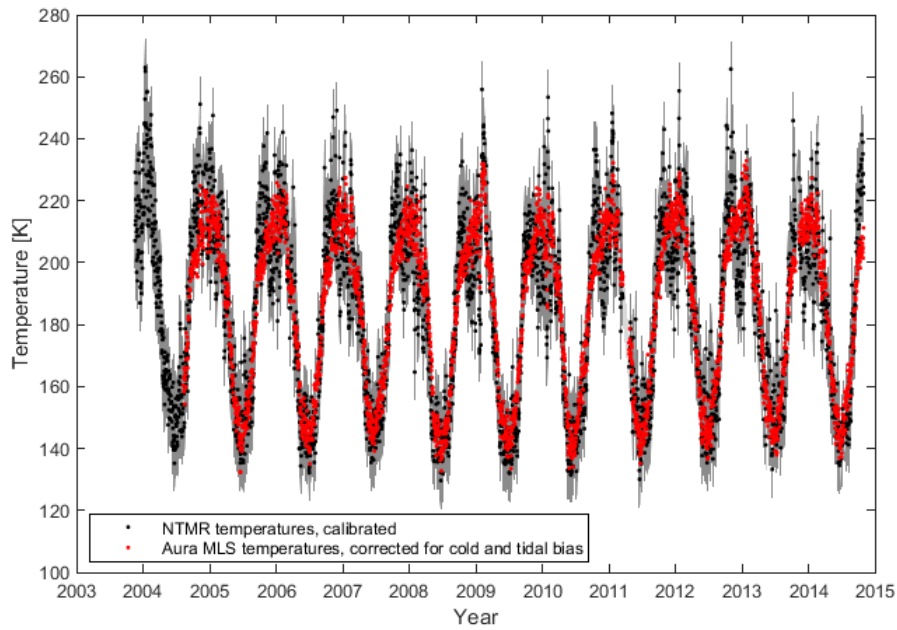


1  
 2 ~~Figure 5. Scatterplot of Aura temperatures corrected for cold and time-of-day bias against~~  
 3 ~~NTMR D<sub>a</sub> rejected temperatures. The red line represents the linear fit to the data.~~

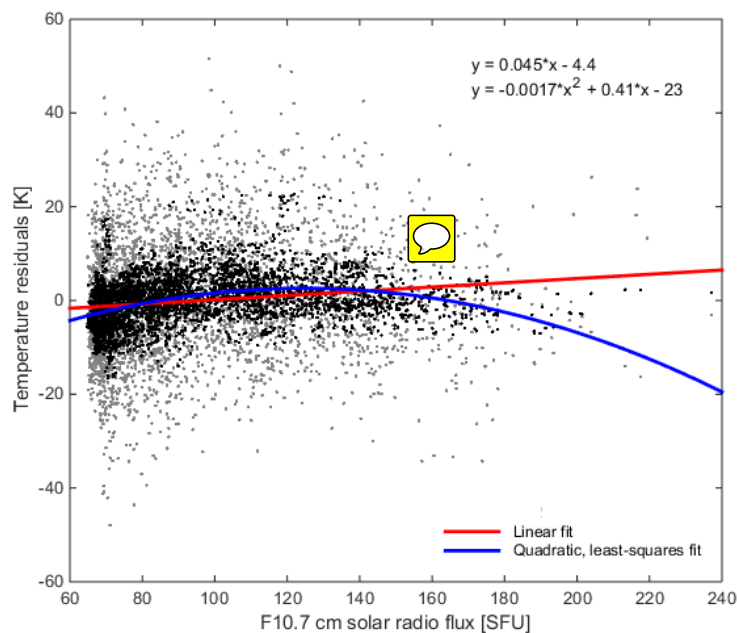


4  
 5 ~~Figure 6. Histogram of the differences between MLS-calibrated NTMR temperatures and~~  
 6 ~~local time and cold bias-corrected Aura MLS temperatures. The red curve is a fitted Gaussian~~  
 7 ~~to the distribution.~~

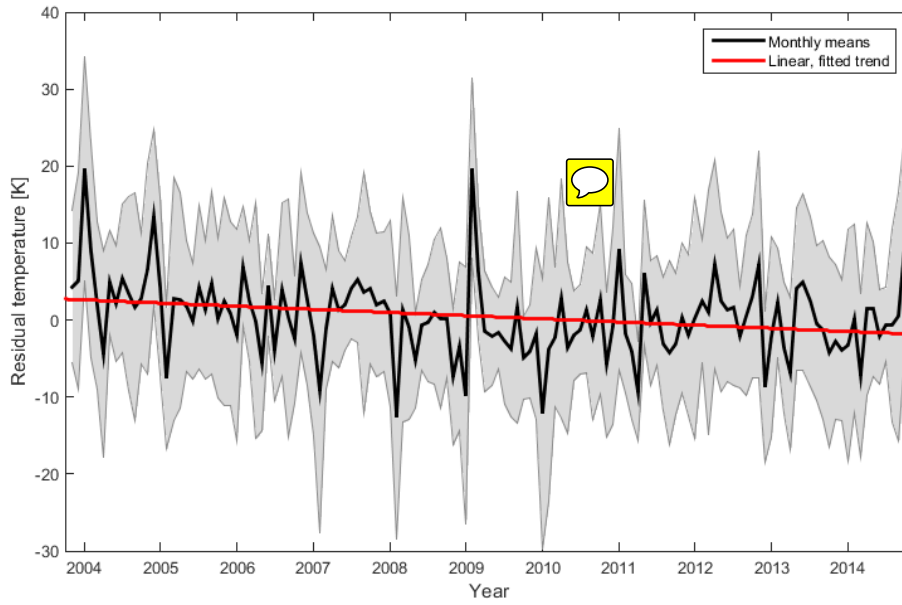




1  
 2 **Figure 7.** Daily values of MLS-calibrated NTMR temperatures plotted together with Aura  
 3 MLS temperatures corrected for cold and time-of-day bias. The overall calibration uncertainty  
 4 is indicated by the grey shading.

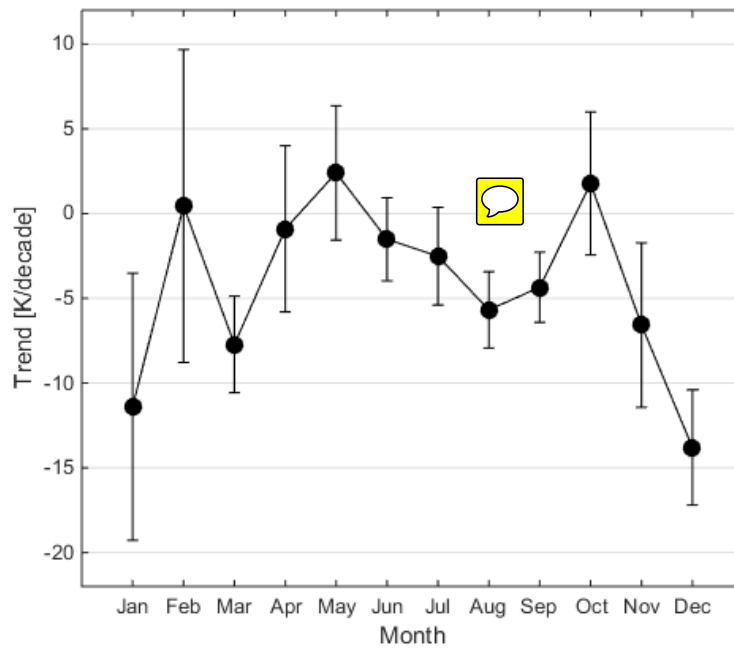


5  
 6 ~~Figure 8. Scatterplot of daily averaged residuals against the corresponding F10.7 cm solar~~  
 7 ~~flux values. Grey dots are daily residuals, while black dots are the residuals with a 30-day~~  
 8 ~~running mean applied. The red line is the linear fit to the daily residuals with the 30-day~~  
 9 ~~running mean applied, while the blue line is the quadratic, least squares fit.~~



1

2 ~~Figure 9. Monthly means of NTMR residual temperatures, corrected for climatology and solar~~  
3 ~~response. The grey shading yields standard deviations.~~



4

5 ~~Figure 10. Monthly long-term temperature trends at 90 km altitude over Tromsø. Standard~~  
6 ~~deviations are given as error bars.~~

