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Change in turbopause altitude at 52 and 70° N

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Abstract

The turbopause is the demarkation between atmospheric mixing by turbulence (below) and molecular diffusion (above). When studying concentrations of trace species in the atmosphere, and particularly long-term change, it may be important to understand processes present, together with their temporal evolution, that may be responsible for redistribution of atmospheric constituents. The general region of transition between turbulent and molecular mixing coincides with the base of the ionosphere, the lower region in which molecular oxygen is dissociated, and, at high latitude in summer, the coldest part of the whole atmosphere.

This study updates previous reports of turbopause altitude, extending the time series by half a decade, and thus shedding new light on the nature of change over solar-cycle timescales. Assuming there is no trend in temperature, at 70° N there is evidence for a summer trend of $\sim 1.2\,\mathrm{km\,decade^{-1}}$, but for winter and at 52° N there is no significant evidence for change at all. If the temperature at 90 km is estimated using meteor trail data, it is possible to estimate a cooling rate, which, if applied to the turbopause altitude estimation, fails to alter the trend significantly irrespective of season.

While studies of atomic oxygen density, [O], using mid-latitude timeseries dating from 1975, show positive trends which can be explained by a lowering of the turbopause, [O] exhibits negative trends since 2002 that, although at a different latitude, are compatible with the observed increase in turbopause height reported here.

1 Introduction

The upper mesosphere and lower thermosphere (UMLT) regime of the atmosphere exhibits a number of features, the underlying physics of which are interlinked and, relative to processes at other altitudes, little understood. At high latitude, the summer mesopause, around 85 km is the coldest region in the entire atmosphere. The UMLT is, inter alia, characterized by the base of the ionosphere, dissociation of molec-

ACPD
15, 20287–20304, 2015

ular species (for example oxygen) by sunlight, and, the focus in this study, the transition from turbulent mixing to distribution of constituents by molecular diffusion. The altitude at which transition turbulence-dominated mixing gives way to molecular diffusion is known as the turbopause, and typically occurs around 100 km, but displaying a seasonal variation, being lower in summer (e.g. ~ 95 km) and higher in winter (e.g. ~ 110 km) (Danilov et al., 1979). Many processes in the UMLT are superimposed and linked. One example is where the mesopause temperature structure determines the altitude dependence of breaking of upwardly propagating gravity waves (e.g. McIntyre, 1991) and thus generation of turbulence. Indeed, the concept of a "wave turbopause" was proposed by Offermann et al. (2007) and compared with the method used forthwith by Hall et al. (2008). Vertical transport by turbulent mixing and horizontal transport by winds redistribute constituents such as atomic oxygen, hydroxyl and ozone. Thus, long-term change in trace constituents cannot be fully explained in isolation from studies of corresponding change in temperature and neutral dynamics.

Determination of the turbopause altitude is non-trivial because direct measurement by radar depends on turbulent structures being "visible" due to small discontinuities in refractive index, e.g. Schlegel et al. (1978) and Briggs (1980). At 100 km, this implies some degree of ionisation and even in situ detectors often depend on ionisation as a tracer (e.g. Thrane et al., 1987). Another means of locating the turbopause is to measure the concentration of particular species as a function of height and noting where the constituents exhibit scale heights that depend on their respective molecular weights, e.g. Danilov et al. (1979). The early work to estimate turbulent energy dissipation rates using medium frequency (MF) radar by Schlegel et al. (1978) and Briggs (1980) was adopted by Hall et al. (1998a). Hall et al. (1998b, 2008) subsequently applied the turbulent intensity estimation to identification of the turbopause. The latter study, which offers a detailed explanation of the analysis, compares methods and definitions and represents the starting point for this study. In addition, Hocking (1983, 1996) and Vandepeer and Hocking (1993) offer a critique on assumptions and pitfalls pertaining to observation of turbulence using radars. One particular pitfall is the problem of group delay of the radar wave in the ionospheric D-region. Of the substantial dataset used in this study, however, only a small percentage of echo profiles are expected to be affected by auroral precipitation that would cause problematic degrees of ionisation below the turbopause. Finally, full descriptions of the radar systems providing the underlying data used here are to be found in Hall (2001) and Manson and Meek (1991).

The characteristic fading time of the signal, τ_c, is used to define an indication of the upper limit for turbulent energy dissipation present in the atmosphere, ε'. First, velocity fluctuations, v' relative to the background wind are identified as:

$$v' = \frac{\lambda\sqrt{\ln 2}}{4\pi\tau_c} \tag{1}$$

where λ is the radar wavelength. This relationship has been presented and discussed by Briggs (1980), Vandepeer and Hocking (1993). In turn v'^2 can be considered to represent the turbulent kinetic energy of the air such that the rate of energy dissipation is obtained by dividing by a characteristic timescale. If the Brunt–Väisälä period T_B ($= 2\pi/\omega_B$ where ω_B is the Brunt–Väisälä frequency in rad s^{-1}) can be a characteristic timescale, then it has been proposed that:

$$\varepsilon' = 0.8v'^2/T_B \tag{2}$$

the factor 0.8 being related to an assumption of a total velocity fluctuation. Alternatively, this can be expressed as:

$$\varepsilon' = 0.8v'^2\omega_B/(2\pi) \tag{3}$$

wherein the Brunt–Väisälä frequency is given by

$$\omega_B = \sqrt{\left(\frac{dT}{dz} + \frac{g}{c_p}\right)\frac{g}{T}} \tag{4}$$

where T is the neutral temperature, z is altitude, g is the acceleration due to gravity and c_p is the specific heat of the air at constant pressure. Due to viscosity, there is a minimum energy dissipation rate, $\varepsilon_{\mathrm{min}}$, present in the atmosphere, given by

$$\varepsilon_{\mathrm{min}} = \omega_{\mathrm{B}}^2 \nu/\beta \tag{5}$$

where ν is the kinematic viscosity. The factor β, known as the mixing or flux coefficient (Oakey, 1982; Fukao et al., 1994; Pardyjak et al., 2002), is related to the flux Richardson Number R_{f} ($\beta = R_{\mathrm{f}}/(1-R_{\mathrm{f}})$). R_{f} is in turn related to the commonly used gradient Richardson number, Ri by the ratio of the momentum to thermal turbulent diffusivities, or turbulent Prandtl number (e.g. Kundu, 1990). Fukao et al. (1994) proposed 0.3 as a value for β. The relationships are fully described by Hall et al. (2008). To use the MF radar system employed here to estimate turbulence is not well suited to estimating Ri due to the height resolution of 3 km; moreover more detailed temperature information would be required to arrive at R_{f}.

Anywhere in the atmosphere, energy dissipation is by the sum of the available processes. In this study, therefore, the turbulent energy dissipation rate can be considered the total rate minus that corresponding to viscosity:

$$\varepsilon = \varepsilon' - \varepsilon_{\mathrm{min}} \tag{6}$$

Importantly, the kinematic viscosity is given by the dynamic viscosity, μ, divided by the density, ρ:

$$\nu = \mu/\rho \tag{7}$$

Thus, since density is inversely proportional to temperature, kinematic viscosity is (approximately) linearly dependent on temperature; ω_{B}^2 is inversely proportional to temperature and therefore $\varepsilon_{\mathrm{min}}$ is approximately independent of temperature. On the other hand, ε' is proportional to ω_{B} and therefore inversely proportional to the square root of temperature.

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

ACPD

15, 20287–20304, 2015

Change in turbopause altitude at 52 and 70° N

C. M. Hall et al.

If we are able to estimate the energy dissipation rates described above, then the turbopause may be identified as the altitude at which $\varepsilon = \varepsilon_{\min}$. This corresponds to equality of inertial and viscous effects and hence the condition where Reynolds number, *Re*, is unity as we shall see.

2 Results and implications for changing neutral air temperature

Following the method described above and by Hall et al. (1998b, 2008), the turbopause position is determined as shown in Fig. 1. The time and height resolutions of the MF radars used for the investigation are 5 min and 3 km respectively, and daily means of turbulent energy dissipation rate profiles are used to determine corresponding turbopause altitudes. The Figure shows the evolution since 1999, 70° N, 19° E (Tromsø) in the upper panel and 52° N, 107° W (Saskatoon) in the lower panel. Data are available from 01 January 1999 to 25 June 2014. The cyan background corresponding to the period 16 February 1999 to 16 October 2000 in the 70° N panel indicates data available but using different experiment parameters. Thus 70° N data prior to 17 October 2000 are excluded from this analysis. A 30 day running mean is shown by the thick lines with the shading either side indicating the standard deviation. The seasonal variation is clear to see, and for illustrative purposes, trend lines have been fitted to June and December values together with hyperbolae showing the 95 % confidence limits in the linear fits (Working and Hotelling, 1929); the seasonal dependence of the trends is addressed in more detail subsequently. It is evident that, apart from the seasonal variation, the mid-latitude turbopause changes little over the period 1999–2014, whereas at high latitude there is more change for the summer state over the period 2001–2014 (the summers of 1999 and 2000 being excluded from the fitting). To investigate the seasonal dependence of the change further, the monthly values for 70 and 52° N are shown in Fig. 2. Since 2001, the high latitude turbopause has increased in height during late spring and mid-summer but otherwise remained constant. The above results represent an update of those by Hall et al. (1998b, 2008), adding more years to the time series. As for the preceding papers and for consistency the neutral atmosphere parameters (temperature and density) required have been obtained from the NRLMSISE-00 empirical model (Picone et al., 2002) and have been assumed not to exhibit any trend over the observation period. In other terms, one-year seasonal climatology temperature models at one-day resolution for 70 and 52° N and altitude range appropriate for the respective radars are therefore used for all years for consistency with earlier results and for consistency between the two latitudes studied here. The only temperature observations available are at 70° N and 90 km altitude and these will be introduced subsequently.

Next, we have attempted to investigate the effects of changing temperature. In a very simplistic approach, hypothetical altitude-invariant trends are imposed on the NRLMSISE-00 profiles. In other words, the same hypothetical trend is applied to all heights in the NRLMSISE-00 profile (for want of better information) to generate evolving (cooling or warming) temperature time-series. The suggested trends vary from −20 to +20 K decade^{-1}, thus well encompassing any realistically conceivable temperature change (cf. Blum and Fricke, 2008; Danilov, 1997; Lübken, 1999). The result of applying hypothetical temperature trends to the time-invariant turbopause heights shown earlier is demonstrated in Fig. 3. Given the seasonal differences identified earlier, four combinations are shown: summer (average of May, June and July) and winter (average of November, December and January) for each latitude. Realistic temperature trends can be considered within the range ±6 K decade^{-1} such that the only significant response of turbopause height to temperature trend is for 70° N in summer. In addition, the figure includes estimated trends obtained from observations, which shall be explained forthwith. The salient point arising from the figure is that no realistic temperature trend (at least given the simple model employed here) has the capability of reversing the corresponding trend in turbopause height.

While a temperature profile covering the UMLT region is not readily available by observations from Tromsø, meteor-trail echo fading times measured by the Nippon/Norway Tromsø Meteor Radar (NTMR) can be used to yield neutral temperatures at 90 km altitude. The method is exactly the same as used by Hall et al. (2012) to deter-

ACPD
15, 20287–20304, 2015

mine 90 km temperatures over Svalbard (78° N) using a radar identical to NTMR. Again, since observational temperature profiles cannot be obtained reliably, NRLMSISE-00 profiles are, of necessity, used in the derivation of turbulent intensity from MF-radar data, but any trend in temperature can usefully be obtained from the meteor radar method (the absolute values of the temperatures being superfluous since they are only available for one height). Hall et al. (2005) investigate the unsuitability of meteor radar data for temperature determination above ~ 95 and below ~ 85 km. In summary: ionization trails from meteors are observed using a radar operating at a frequency less than the plasma frequency of the electron density in the trail (this is the so-called "underdense" condition). It is then possible to derive ambipolar diffusion coefficients D from the radar echo decay times τ according to:

$$\tau = \frac{\lambda^2}{16\pi^2 D} \tag{8}$$

wherein λ is the radar wavelength. Thereafter the temperature T may be derived using the relation:

$$T = \sqrt{\frac{P \cdot D}{6.39 \times 10^{-2} K_0}} \tag{9}$$

where P is the neutral pressure and K_0 is the zero field mobility of the ions in the trail (here we assume $K_0 = 2.4 \times 10^{-4}\,\mathrm{m}^{-2}\,\mathrm{s}^{-1}\,\mathrm{V}^{-1}$) (McKinley, 1961; Chilson et al., 1996; Cervera and Reid, 2000; Holdsworth et al., 2006). The pressure, P, can be obtained from NRLMSISE-00 for consistency with the turbulence calculations.

In a recent study, Holmen et al. (2015) have built on the method of Hall et al. (2012) to determine 90 km temperatures over NTMR. This new work includes more sophisticated approaches to normalisation to independent measurements and to investigating the dependence of derived temperatures on solar flux. Having removed seasonal and solar cycle variations in order to facilitate trend-line fitting (as opposed to isolating a hypothetical anthropogenic-driven variation), Holmen et al. (2015) arrive at a temperature

ACPD
15, 20287–20304, 2015

trend of $-3.6 \pm 1.1\,\mathrm{K\,decade^{-1}}$ determined over the time interval 2004–2014 inclusive. This can be considered statistically significant (viz. significantly non-zero at the 5 % level) since the uncertainty ($2\sigma = 2.2\,\mathrm{K\,decade^{-1}}$) is less than the trend itself (e.g. Tiau et al., 1990).

Estimation of changes in temperature corresponding to the period for determination of the turbopause were only viable for 70° N, these being $-0.8 \pm 2.9\,\mathrm{K\,decade^{-1}}$ for summer and $-8.1 \pm 2.5\,\mathrm{K\,decade^{-1}}$ for winter, and these results are indicated in Fig. 3. Again using the simple idea of superimposing a gradual temperature change (the same for all heights) on the temperature model used for the turbulence determination thus fails to alter the change in turbopause height significantly, for the ~ decade of observations. Although direct temperature measurements are not available for the 52° N site, Offermann et al. (2010) report cooling rates of $\sim 2.3\,\mathrm{K\,decade^{-1}}$ for 51° N, 7° E, and She et al. (2015) $\sim 2.8\,\mathrm{K\,decade^{-1}}$ for 42° N, 112° W. As for 70° N, these results do not alter the conclusions inferred from Fig. 3.

3 Discussion

The aim of this study has been to update earlier reports (viz. Hall et al., 2008) of turbopause altitude and change determined for two geographic locations: 70° N, 19° E (Tromsø) and 52° N, 107° W (Saskatoon). An effort has been made to demonstrate that conceivable temperature trends are unable to alter the overall results, viz. that there is evidence of increasing turbopause altitude at 70° N, 19° E in summer, but otherwise no significant change during the period 2001 to 2014. Assimilating results from in situ experiments spanning the time interval 1966–1992, Pokhunkov et al. (2009) present estimates of turbopause height trends for several geographical locations, but during a period prior to that of our observations. For high latitude the turbopause is reported to have fallen by ~ 2–4 km between 1968 and 1989 – the opposite sign of our finding for 2001–2014. More recently, further evidence has been presented for a long-term descent of the turbopause, at least at mid-latitude (Oliver et al., 2014 and references

ACPD
15, 20287–20304, 2015

therein). The rationale for this is that the atomic oxygen density [O] has been observed to increase during the time interval 1975–2014 at a rate of approximately 1 % year^{-1}. The associated change in turbopause height may be estimated thus:

$$H = RT/(mg) \tag{10}$$

where H is scale height, R is the universal gas constant (= 8.314 J mol^{-1} K^{-1}), m is the mean molecular mass (kg mol^{-1}) and g is the acceleration due to gravity. At 120 km altitude, g is taken to be 9.5 m s^{-2}. For air and atomic oxygen, m = 29 and 16 respectively. For a typical temperature of 200 K, the two corresponding scale heights are therefore H_{air} = 6.04 km and H_{oxygen} = 10.94 km. If the change (fall) in turbopause height is denoted by Δh_{turb}, then Oliver et al. (2014) indicate that the factor by which [O] would increase is given by:

$$\exp(\Delta h_{\text{turb}}/H_{\text{air}})/\exp(\Delta h_{\text{turb}}/H_{\text{oxygen}}) \tag{11}$$

Note that Oliver et al. (2014) state that "[O]... would increase by the amount", but, since Eq. (11) is dimensionless, the reader should be aware this is a factor, not an absolute quantity. At first, there would appear to be a fundamental difference between the findings derived from [O] at a mid-latitude station and those for ε from a high-latitude station, and indeed the paradox could be explained by either the respective methods and/or geographic locations. However if one examines the period from 2002 onwards (corresponding to the high-latitude dataset, but only about one quarter of that from the mid-latitude station), a decrease in [O] corresponds with an increase in Δh_{turb}. If, then, Δh_{turb} for the measured summer temperature change at high latitude (viz. 0.18 km year^{-1} from Fig. 3) is inserted in Eq. (11) together with the suggested scale heights for air and atomic oxygen, one obtains a corresponding decrease in [O] of 13 % decade^{-1}, e.g. over the period 2002–2012. The corresponding time interval is not analysed per se by Oliver et al. (2014) but a visual inspection suggests a decrease of the order of 20 %; the decrease itself is incontrovertible and therefore in qualitative agreement with our high-latitude result.

ACPD
15, 20287–20304, 2015

It is somewhat unfortunate that it is difficult to locate simultaneous and approximately co-located measurements by different methods. The turbopause height-change derived by Oliver et al. (2014) are by measurements of [O] and at mid-latitude; those by Pokhunkov et al. (2009), also by examining constituent scale-heights, include determinations for Heiss Island (80° N, 58° E) but this rocket sounding programme was terminated prior to the start of our observation series (Danilov et al., 1979). It should be noted, however that the results of seasonal variability presented by Danilov et al. (1979) agree well with those described here giving credence to the method and to the validity of the comparisons above.

4 Conclusion

Updated temporal evolutions of the turbopause altitude have been presented for two locations: 70° N, 19° E (Tromsø) and 52° N, 107° W (Saskatoon), the time interval now spanning 1999 to 2014. These turbopause altitude estimates are derived from estimates of turbulent energy dissipation rate obtained from medium-frequency radars. The method entails a knowledge of neutral temperature which, earlier (Hall et al., 2008) had been assumed to be constant with time. Here the response of the change in turbopause heights over the period of the study to temperature trends – both hypothetical and observed – is examined. No temperature trend scenario was capable of altering the observed turbopause characteristics significantly; at 70° N an increase in turbopause height is evident during the 1999–2014 period for summer months, whereas for winter at 70° N and all seasons at 52° N the turbopause height has not changed.

At first, this conclusion would appear to contradict the recent report by Oliver et al. (2014) and Pokhunkov et al. (2009), however, closer inspection shows that if one considers the time interval 2002–2012 in isolation, there is a qualitative agreement. In fact we note that Oliver et al. (2014) deduce a turbopause change based on changing atomic oxygen concentration and so we are able to similarly deduce a change in atomic oxygen concentration based on the change in turbopause height obtained from direct estimation of turbulence intensity. Given an average (i.e. not differentiating between seasons) temperature change of $-3.4 \pm 0.5\,\mathrm{K\,decade^{-1}}$ for 70° N, 19° E (Tromsø), the change in turbopause height in summer over the same time interval is $1.7 \pm 0.2\,\mathrm{km\,decade^{-1}}$ suggesting a decrease in atomic oxygen concentration of 13 %.

The primary result of this study is to demonstrate the increasing altitude of the summer turbopause at 70° N, 19° E and the apparently unvarying altitude in winter and at 52° N, 107° W during the time interval 1999–2014. Independent studies using a radically different method demonstrate how to infer a corresponding decrease in atomic oxygen concentration, as a spin-off result. Finally, the question as to the exact mechanism causing the evolution of turbulence in the lower thermosphere at, in particular 70° N, 19° E, remains unanswered, and furthermore, dynamics at this particular geographic location may be pathological. The solution perhaps lies in seasonally dependent gravity wave filtering in the underlying atmosphere being affected by climatic tropospheric warming and/or middle atmosphere cooling; hitherto, however, this remains a hypothesis.

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ACPD
15, 20287–20304, 2015

Change in turbopause altitude at 52 and 70° N

C. M. Hall et al.

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀| |▶
◀ ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

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ACPD
15, 20287–20304, 2015

Change in turbopause altitude at 52 and 70° N

C. M. Hall et al.

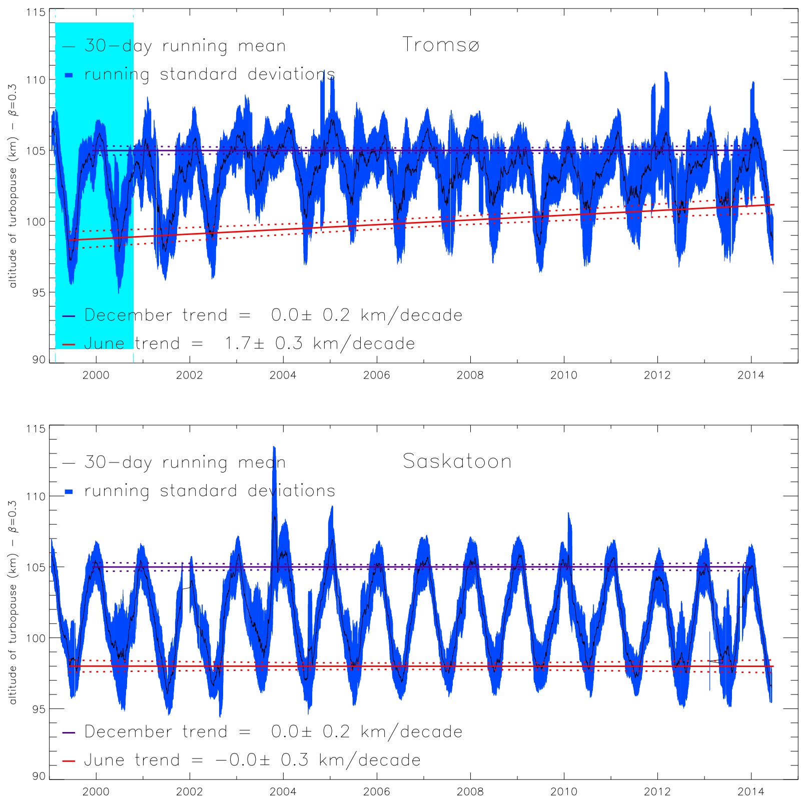

Figure 1. Turbopause altitude as determined by the definition and method described in this paper. The thick solid line shows the 30 day running mean and the shading behind it the corresponding standard deviations. The straight lines show the fits to summer and winter portions of the curve. Upper panel: 70° N (Tromsø); lower panel: 52° N (Saskatoon). The cyan background in the 70° N panel indicates data available but unused here due to different experiment parameters.

ACPD
15, 20287–20304, 2015

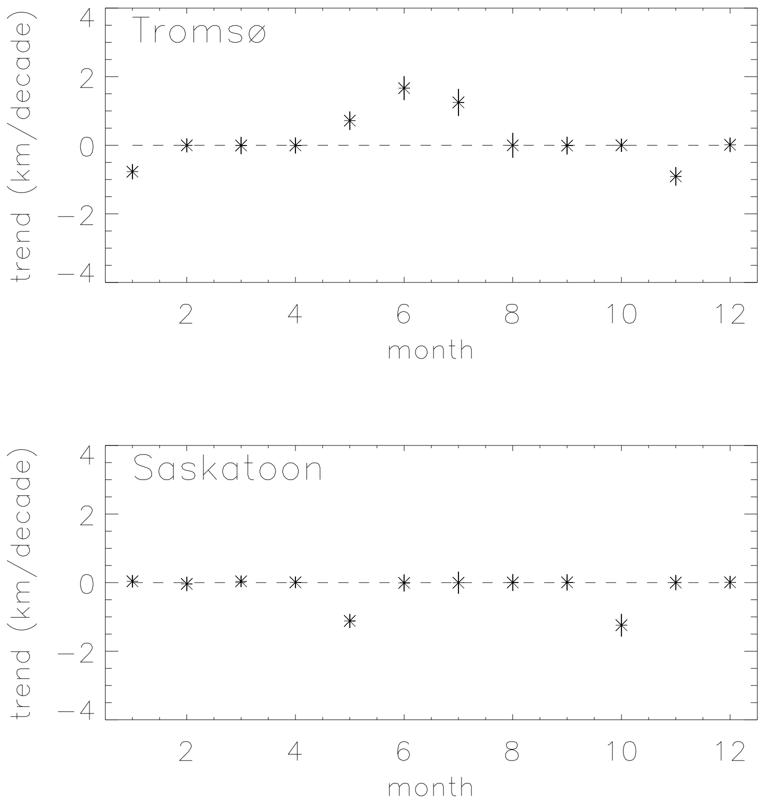

Figure 2. Trends for the period as a function of month. Upper panel: 70° N (Tromsø); lower panel: 52° N (Saskatoon).

ACPD
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Change in turbopause altitude at 52 and 70° N

C. M. Hall et al.

Title Page
Abstract Introduction
Conclusions References
Tables Figures
◂ ▸
Back Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

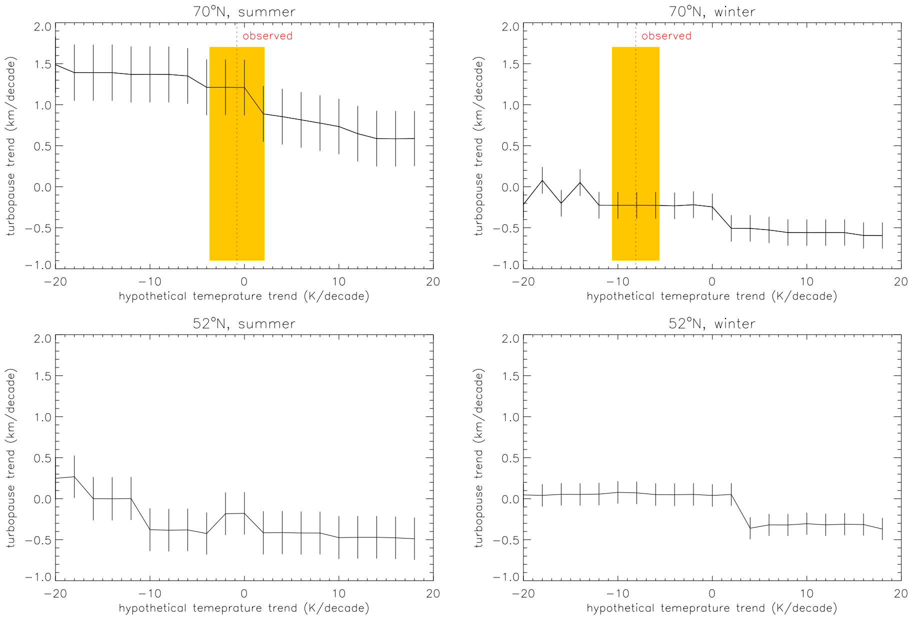

Figure 3. Response of turbopause trend line to different upper-mesosphere/lower-thermosphere temperature trends. Hypothetical trends range from an unrealistic cooling of $20\,\mathrm{K\,decade^{-1}}$ to a similarly unrealistic warming. Top-left: 70° N summer (average of May, June and July); top-right: 70° N winter (average of November, December and January); bottom-left: 52° N summer; bottom-right: 52° N winter. Observed values for 70° N are also identified on the upper panels (dashed vertical lines) together with uncertainties (shading).