Change in turbopause altitude at 52° and 70°N

C. M. Hall[1], S. E. Holmen[1, 2, 5], C. E. Meek[3], A. H. Manson[3] and S. Nozawa[4]

[1] Tromsø Geophysical Observatory, UiT - The Arctic University of Norway, Tromsø, Norway

[2] The University Centre in Svalbard, Norway

[3] University of Saskatchewan, Saskatoon, Canada

[4] Nagoya University, Nagoya, Japan

[5] Birkeland Centre for Space Science, Bergen Norway

Correspondence to:

C. M. Hall (chris.hall@tgo.uit.no)

Full institute address

UiT - The Arctic University of Norway,

Tromsø Geophysical Observatory,

9037 Tromsø,

Norway

Abstract

The turbopause is the demarcation 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 ~1.6 km/decade, 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.

The observed increase in turbopause height supports a hypothesis of corresponding negative trends in atomic oxygen density, [O]. This supports independent studies of atomic oxygen density, [O], using mid-latitude timeseries dating from 1975, which show negative trends since 2002..

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 85km is the coldest region in the entire atmosphere. The UMLT is, inter alia, characterized by the base of the ionosphere, dissociation of molecular 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. ~110km) (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). Prevailing winds filter or even inhibit propagation of gravity waves generated in the lower atmosphere and the static stability (or lack of it) of the atmosphere dictates the vertical distribution of gravity wave saturation and breaking. The generation of turbulence and its height distribution vary with season and similarly affect the turbopause altitude (e.g. Hall et al., 1997). Turbulence is somewhat distributed through the high latitude winter mesosphere, whereas in summer the gravity waves "save their energy" more until reaching the "steep beach" (a visualization attributable to M. E. McIntyre - private communication) of the summer mesopause near 85km. 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.

One 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). Detection of turbulence and estimation of its intensity 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 100km, this implies some degree of ionisation and even in situ detectors often depend on ionisation as a tracer (e.g. Thrane et al. 1987). A common means of quantifying turbulent intensity is the estimation of turbulent energy dissipation rate, ε. In the classical visualization of turbulence in two dimensions, large vortices generated by, for example breaking gravity waves or wind shears form progressively smaller vortices (eddies) until inertia is insufficient to overcome viscous drag in the fluid. Viscosity then "removes" kinetic energy and transforms it to heat. This "cascade" from large-scale vortices to the smallest scale eddies capable of being supported by the fluid, and subsequent dissipation of energy, was proposed by Kolmogorov (1941) but more accessibly described by Batchelor (1953) and (e.g.) Kundu (1990). At the same time, a minimum rate of energy dissipation by viscosity is supported by the atmosphere (defined subsequently). The altitude at which these two energy dissipation rates are equal is also a definition of the turbopause and corresponds to the condition where the Reynolds number, the ratio between inertial and viscous forces, is unity.

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). The reader is referred to these earlier publications for a full explanation, but in essence, velocity fluctuations relative to the background wind give rise to fading with time of echoes from structures in electron density drifting through the radar beam. While the drift is determined by cross-correlation of signals from spaced receiver antennas, autocorrelation yields fading times which may be interpreted as velocity fluctuations (the derivation of which is given in the following section). The squares of the velocity perturbations can be equated to turbulent kinetic energy and then, when divided by a characteristic timescale become energy dissipation rates. Energy is conserved in the cascade to progressively smaller and more numerous eddies such that the energy dissipation rate is representative of the ultimate conversion of kinetic energy to heat by viscosity. Hall et al. (1998b and 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 and 1996) and Vandepeer and Hocking (1993) offer a critique on assumptions and pitfalls pertaining to observation of turbulence using radars. For the radars to obtain echoes from the UMLT, a certain degree of ionization must be present and daylight conditions yield better results than night-time, and similarly results are affected by solar cycle variation. However, there is a trade-off: too little ionization prevents good echoes while too much gives rise to the problem of group delay of the radar wave in the ionospheric D-region. Space weather effects that are capable of creating significant ionization in the upper mesosphere are infrequent, and aurora normally occur on occasional evenings at high latitude, and then only for a few hours duration at the most. 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. While it must be accepted that group delay at the radar frequencies used for the observations reported here cannot be dismissed, the MF-radar method is the only one that has been available for virtually uninterrupted measurement of turbulence in the UMLT region over the past decades.

Full descriptions of the radar systems providing the underlying data used here are to be found in Hall (2001) and Manson and Meek (1991) and the salient features of the radars, relevant for this study are given in Table 1.

Analysis methodology

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', as explained above. First, velocity fluctuations, v´ relative to the background wind are identified as:

$$\mathrm{v}' = \frac{\lambda\sqrt{\ln 2}}{4\pi\,\tau_c} \qquad (1)$$

where λ is the radar wavelength. This relationship has been presented and discussed by Briggs (1980) and 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 dissipation of this energy 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:

$$\varepsilon' = 0.8\mathrm{v}'^2 / T_B \qquad (2)$$

the factor 0.8 being related to an assumption of a total velocity fluctuation (see Weinstock, 1978). Alternatively, this can be expressed as:

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

wherein the Brunt-Väisälä frequency is given by

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

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, $\varepsilon_{\min}$, present in the atmosphere, given by

$$\varepsilon_{\min} = \omega_B{}^2 \nu \big/ \beta \qquad (5)$$

where ν is the kinematic viscosity. The factor β, known as the mixing or flux coefficient (Oakey, 1982; Fukao et al., 1994; Pardyjac et al., 2002), is related to the flux Richarson Number R_f ($\beta = R_f/(1\text{-}R_f)$). 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 3km; 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:

$$\varepsilon = \varepsilon' - \varepsilon_{\min} \qquad (6)$$

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

$$\nu = \mu \big/ \rho \qquad (7)$$

Thus, since density is inversely proportional to temperature, kinematic viscosity is (approximately) linearly dependent on temperature; $\omega_B{}^2$ is inversely proportional to temperature and therefore $\varepsilon_{\min}$ 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.

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

To implement the above methodology, temperature data are required. Since observational temperature profiles cannot be obtained reliably, NRLMSISE-00 empirical model (Picone et al., 2002) profiles are, of necessity, used in the derivation of turbulent intensity from MF-radar data. The reasons for this are discussed in detail in the following section. While a temperature profile covering the UMLT region is not readily available by ground-based 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. Any trend in temperature can usefully be obtained (the absolute values of the temperatures being superfluous since they are only available for one height). The method is exactly the same as used by Hall et al. (2012) to determine 90 km temperatures over Svalbard (78°N) using a radar identical to NTMR. Hall et al. (2005) investigate the unsuitability of meteor radar data for temperature determination above ~95km 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_{meteor}, (as distinct from the corresponding fading time for the medium-frequency radars) according to:

$$\tau_{meteor} = \frac{\lambda^2}{16\pi^2 D} \qquad (8)$$

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

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

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 = 2.4×10^{-4} m^{-2} s^{-1} V^{-1}) (McKinley, 1961; Chilson et al., 1996; Cervera and Reid, 2000; Holdsworth et al., 2006). The pressure, P, was obtained from NRLMSISE-00 for consistency with the turbulence calculations. In the derivations by Dyrland et al. (2010) and Hall et al. (2012), for example, temperatures were then normalized to independent measurements by the MLS (Microwave Limb Sounder) on board the EOS (Earth Observing System) Aura spacecraft launched in 2004. The MLS measurements were chosen because the diurnal coverage was constant for all measurements and it was therefore simpler to estimate values that were representative of daily means, than other sources such as SABER. In this way, the influence of any systematic deficiencies in NRLMSISE-00 (e.g. due to the age of the model) were minimized.

Results and implications for changing neutral air temperature

Following the method described above and by Hall et al. (1998b and 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 minutes 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. Results are, of course specific to these geographical locations and it must be stressed that they are in no way zonally representative (hereafter, though, "70°N" and "52°N" may be used to refer to the two locations for convenience). Data are available from 1 January 1999 to 25 June 2014 for Saskatoon but thereafter, technical problems affected data quality. Data are shown from 1 January 1999 to 25 October 2015 for

Tromsø. The cyan background corresponding to the period 16 February 1999 to 16 October 2000 in the 70°N (Tromsø) panel indicates data available but using different experiment parameters and 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. The months of June and December are chosen simply because these correspond to the solstices and thus to avoid any a priori conception of when one could anticipate the maxima and minima to be. 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 2001-2015 (the summers of 1999 and 2000 being excluded from the fitting due to changes in experiment parameters for the Tromsø radar). To investigate the seasonal dependence of the change further, the monthly values for 70°N 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. Since individual months are selected the possibility of "end-point" biases are not an issue in the trend-line fitting as would be the case if analyzing entire datasets with non-integer numbers of years. Even so, certain years may be apparently anomalous, for example the summer of 2003. In this study, the philosophy is to look for any significant change in the atmosphere over the observational period. If anomalous years are caused by, for example, changes in gravity-wave production (perhaps due to an increasing frequency of storm in the troposphere) and filtering in the underlying atmosphere, these too should be considered part of climate change. The trend (or overall change) over the observation period is indeed sensitive to exclusion of certain years. Although not illustrated here, this was tested briefly: selecting data from only 2004 onwards indicates no significant change for summer, but a slightly increased negative winter change (to -1.7 ± 0.2 K $^{\text{decade-1}}$); excluding only 2003 (the visually anomalous year) fails to alter the summer and winter values significantly at all. The above findings represent an update of those by Hall et al. (1998b and 2008), adding more years to the time series and therefore now covering a little over one solar cycle (the latter half of cycle 23 and first half of 24). 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°N 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. Satellite-based temperature determinations are, of course available, including, for example SABER (Sounding of the Atmosphere by Broadband Emission Radiometry) on board TIMED (Thermosphere Ionosphere Mesosphere Energetics and Dynamics) which was launched in 2001. The temporal sampling by such instruments makes the estimation of (for example) daily means somewhat complicated. Moreover, the measurements are not necessarily representative for the field of field of view of the radar because the geographical coverage of remote sensing data needs to be sufficiently large to obtain the required annual coverage, since the sampling region can vary with season (depending on the satellite). Choice of the somewhat dated NRLMSISE-00 model at least allows the geographical location to be specified and furthermore ensures a degree of consistency between the two sets of radar observations and also earlier analyses. The only ground-based temperature observations both available and suitable are at 70°N and 90 km altitude as described earlier and used 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 (for want of better information) in the NRLMSISE-00 profile to generate evolving (cooling or warming) temperature time-series. The suggested trends vary from -20Kdecade^{-1} to +20Kdecade^{-1}, thus well encompassing any realistically conceivable temperature change (c.f. 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 Figure 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 geographic location. Realistic temperature trends can be considered within the range ±6 Kdecade^{-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.

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, as has been described in the previous section. This new work presents more sophisticated approaches for normalisation to independent measurements and 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 trend of -3.6 ± 1.1 Kdecade^{-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 (2σ = 2.2 Kdecade^{-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 ± 2.9 Kdecade^{-1} for summer and -8.1 ± 2.5 Kdecade^{-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 ~2.3 K decade^{-1} for 51°N, 7°E, and She et al. (2015) ~2.8 K decade^{-1} for 42°N, 112°W. As for 70°N, these results do not alter the conclusions inferred from Fig. 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 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 ms^{-2}. For air and atomic oxygen, m = 29 and 16 respectively. For a typical temperature of 200K, 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:

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

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. Usefully, in this context, Shinbori et al. (2014) and Kozubek et al. (2015) investigate such geographical diversity. 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. 0.16 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 16% decade^{-1}, e.g. over the period 2002-2015. 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.

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.

Finally, the change in turbopause altitude during the last decade or more should be placed in the context of other observations. The terrestrial climate is primarily driven by solar forcing, but several solar cycles of data would be required to evaluate the effects of long-term change in space weather conditions on turbulence in the upper atmosphere. A number of case-studies have been reported, however that indicate how space weather events affect the middle atmosphere (Jackman et al., 2005; Krivolutsky et al., 2006). One recurring mechanism is forced change in stratospheric chemistry (in particular, destruction and production of ozone and hydroxyl); the associated perturbations in temperature structure adjust the static stability of the atmosphere through which gravity waves propagate before reaching the mesosphere. In addition, greenhouse gases causing global warming in the troposphere act as refrigerants in the middle atmosphere and so changing the static stability and therefore the degree to which gravity waves shed turbulence en route to the UMLT. Not a subject of this study, it is hypothesised that changes in the troposphere and oceans give rise to a higher frequency of violent weather; this in turn could be expected to increase the overall gravity wave activity originating in the lower atmosphere but propagating through the middle atmosphere. Sudden stratospheric warmings (SSWs) also affect (by definition) the vertical temperature structure and thus gravity wave propagation (e.g. de Wit et al., 2015; Cullens et al., 2015). Apart from direct enhancements of stratospheric temperatures, SSWs have been demonstrated to affect planetary wave activity even extending into the opposite hemisphere (e.g. Stray et al., 2015). If such effects were capable of, for example, triggering the springtime breakdown of the polar vortex, associated horizontal transport of stratospheric ozone contributes to determination of the tropopause altitude (e.g. Hall, 2013) and again, gravity wave propagation. Overall change in the stratosphere is proposed as the origin of the observed strengthening of the Brewer Dobson circulation during the last 35 years at least (Fu et al., 2015). Closer to the 70°N, 19°E (Tromsø) observations, Hoffmann et al. (2011) report increases in gravity wave activity at 55°N, 13°E during summer, including at 88km. Although not co-located, the increasing gravity wave flux, with waves breaking at the summer high latitude mesopause would similarly increase turbulence intensity and support the change reported here. Further references to long-term change in the middle and upper atmosphere in general can be found in Cnossen et al. (2015). Background winds and superimposed tides thus affecting gravity wave propagation and filtering in the atmosphere underlying the UMLT also vary from location to location at high latitude and the two studies by Manson et al. (2011a and 2011b) study this zonal difference and compare with a current model. Although for approximately 10° further north than the Tromsø radar site, these studies give valuable background information, on not only the wind field, but also on tidal amplitude perturbation due to deposition of gravity waves' horizontal momentum.

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 2015. 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 that had earlier (Hall et al., 2008) 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, 19°E an increase in turbopause height is evident during the 1999-2015 period for summer months, whereas for winter at 70°N, 19°E and all seasons at 52°N, 107°W the turbopause height has not changed significantly. In evaluating these results, however, there are a number of caveats that must be remembered. Firstly, the radar system does not perform well with an aurorally disturbed D-region – the study, on the other hand incorporates well over 100,000 hours of data for each radar site, and auroral conditions are occasional and of the order of a few hours each week at most. Secondly, an influence of the semi-empirical model used to provide both density and Brunt-Väisälä frequencies cannot be disregarded. It should also be stressed that a change is being reported for the observational periods of approximately 15 years (i.e. just over one solar cycle) and parameterized by fitting linear trend-lines to the data; this is distinct from asserting long-term trends in which solar and anthropogenic effects can be discriminated.

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 similarly able to 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±0.5 K decade^{-1} for 70°N, 19°E (Tromsø), the change in turbopause height in summer over the same time interval is 1.6±0.3 km decade^{-1} suggesting a decrease in atomic oxygen concentration of 16%.

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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Table 1. Salient radar parameters

Parameter	**Tromsø**	**Saskatoon**
Geographic coordinates	69.58°N, 19.22°E	52.21°N, 107.11°E
Operating frequency	2.78 MHz	2.22 MHz
Pulse length	20 μs	20 μs
Pulse repetition frequency	100 Hz	60 Hz
Power (peak)	50 kW	25 kW
Antenna beamwidth	17° at -3dB	17° at -6dB
Altitude resolution	3 km	3 km
Time resolution (post-analysis	5 min	5 min

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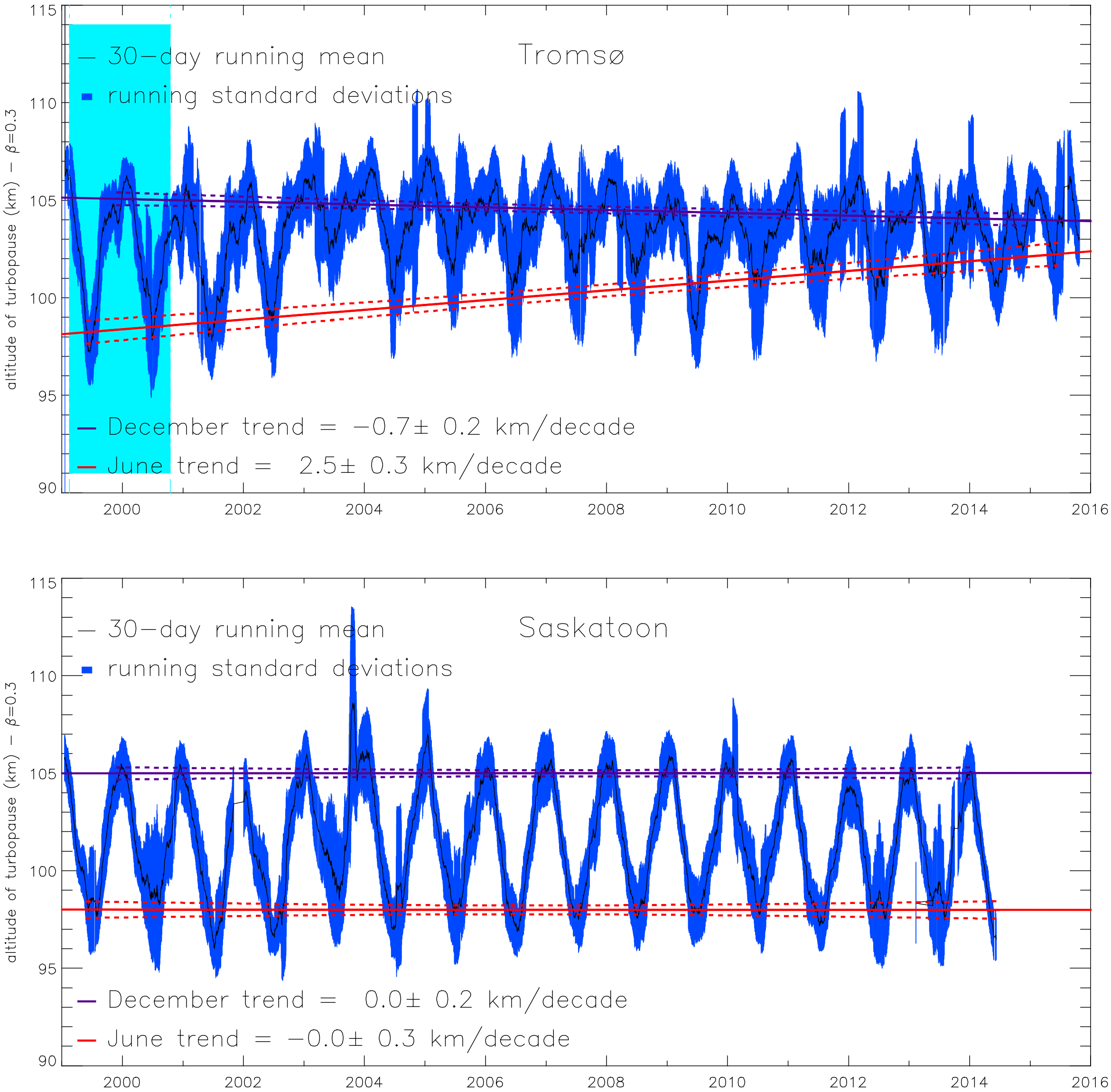

Figure 1. Turbopause altitude as determined by the definition and method described in this paper. The thick solid line shows the 30-day running mean and the shading behind it the corresponding standard deviations. The straight lines show the fits to summer and winter portions of the curve. Upper panel: 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

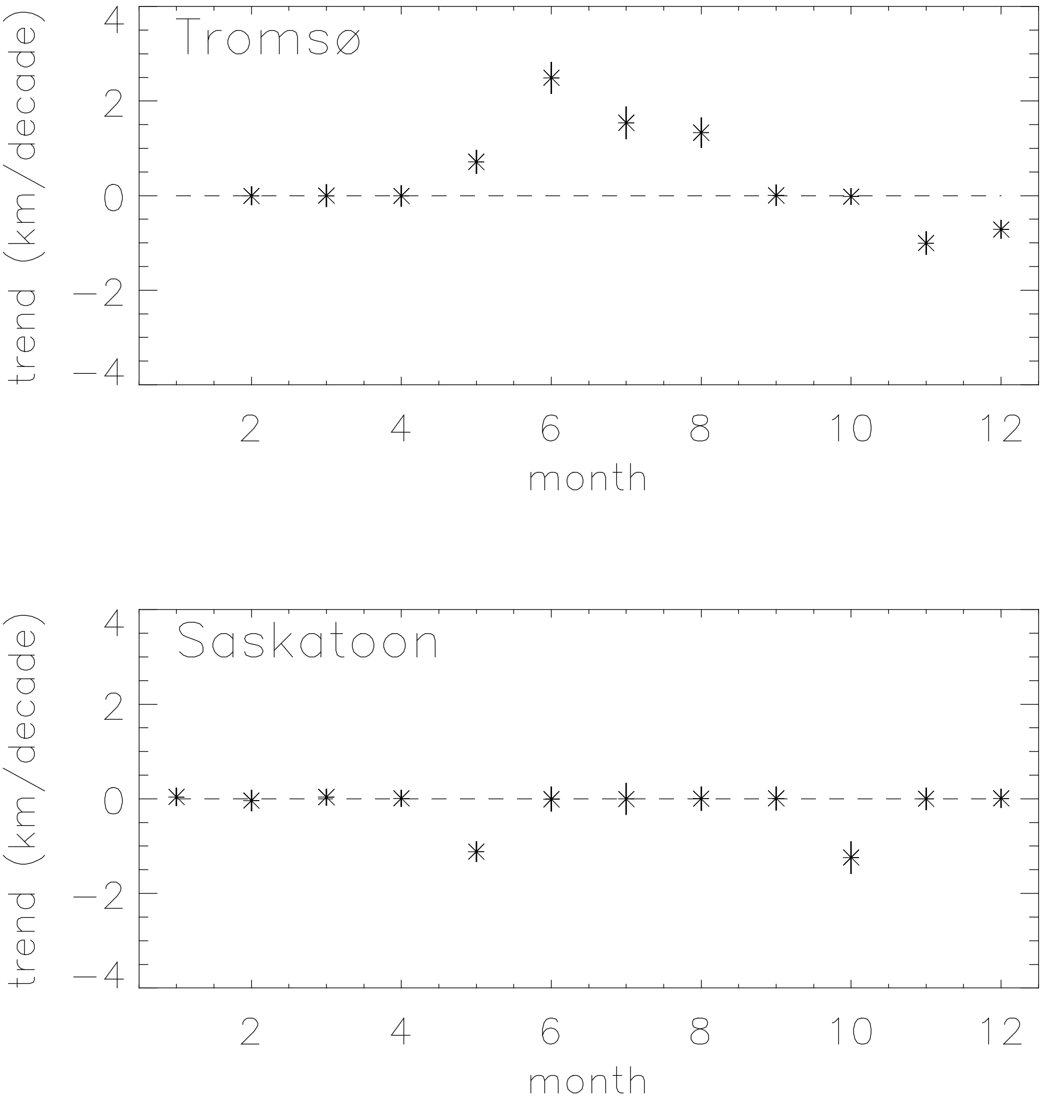

Figure 2. Trends for the period as a function of month. . Upper panel: 70°N (Tromsø); lower panel: 52°N (Saskatoon).

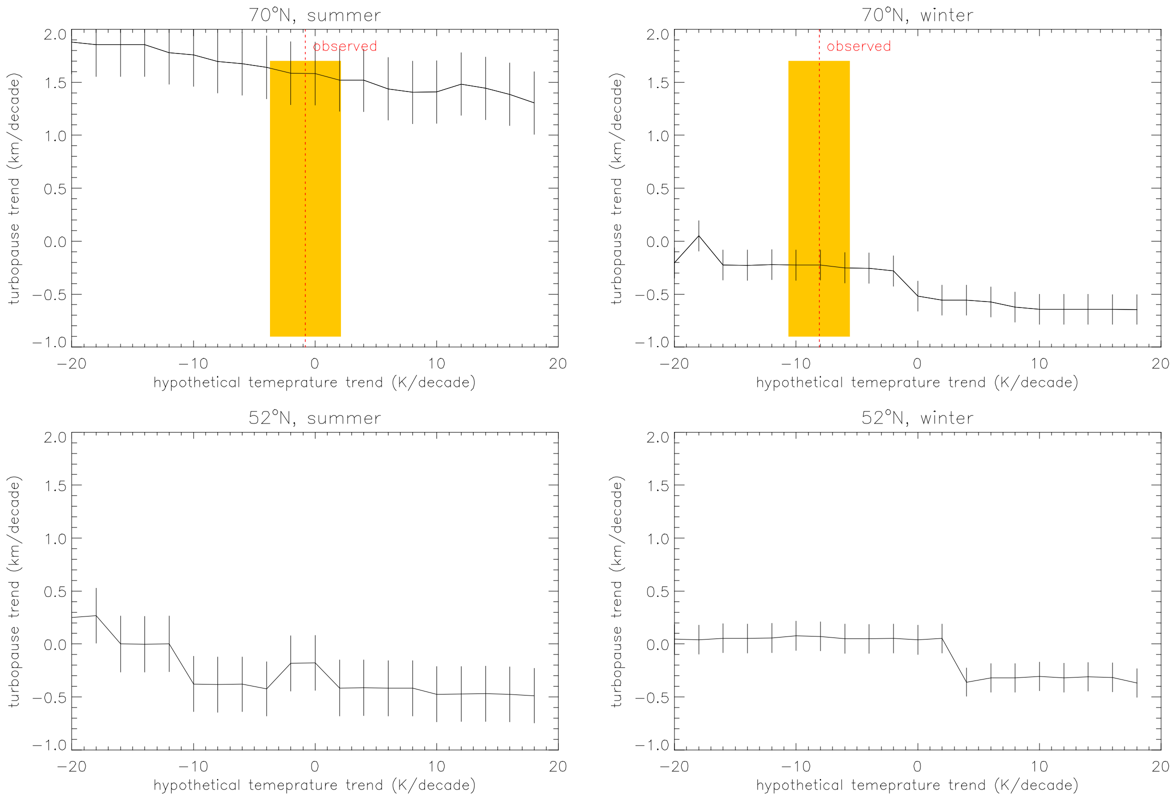

Figure 3. Response of turbopause trend line to different upper-mesosphere/lower-thermosphere temperature trends. Hypothetical trends range from an unrealistic cooling of 20K/decade to a similarly unrealistic warming. Top-left: 70°N summer (average of May, June and July); top-right: 70°N winter (average 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).