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The Hydroxyl Rotational Temperature Record From the Auroral Station in Adventdalen, Svalbard (78°N, 15°E): Preliminary Results.

F. Sigernes[1], K. P. Nielsen[2], C. S. Deehr[3], T. Svenøe[1], N. Shumilov[1,4], and O. Havnes[1,4]

[1]The University Courses on Svalbard, N-9171 Longyearbyen, Norway.
[2]The University Courses on Svalbard and the University of Bergen, Norway
[3]Geophysical Institute, University of Alaska, Fairbanks, USA.
[4]The Auroral Observatory, University of Tromsø, Norway

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Abstract. This paper describes the compiled mesospheric winter temperature series derived through over 20 years of ground– based spectral measurements of the hydroxyl airglow layer from the Auroral Station in Adventdalen near Longyearbyen, Svalbard (78°N, 15°E). Diurnal and semi diurnal tides are present. Temperature changes of 20°K with extremes up to 70°K within a few hours are usual. The average daily winter temperature for the period was found to be 208°K with a standard deviation of 16°K. The temperature series is interspersed with periods of 2–5 days when the temperatures may drop down to the summer minimum. These minima are probably associated with stratospheric warmings that occur with varying strength every winter. Some years are more quiet than others with respect to variation in temperature. The December-January dip in the temperatures around New year is not clearly identified for all seasons. The annual mean winter temperatures show nearly zero temperature trend, indicating no statistically significant change in the winter mesospheric temperatures over Svalbard during the last two decades.

1 Introduction

The mesosphere and lower thermosphere may be the least known parts of our atmosphere, partly because of the problem of obtaining direct in situ observations. On the other hand it appears that this part of the atmosphere may be most, and first, affected by changes in the atmospheric content of greenhouse gases. There are clear signs that these parts of the atmosphere have changed profoundly at least during the last three decades.

The natural thermal structure is believed to be primarily controlled by heating from absorption of short wave solar radiation and as a response balanced by cooling related to emissions in the infrared part of the spectrum. The absorption of UV by ozone (O_3) constitutes the principal radiative source of heat in the stratosphere and mesosphere. In addition, absorption of UV by molecular oxygen (O_2) plays an important role, especially in the upper mesosphere and lower thermosphere. The absorption processes occur as dissociation, forming atomic oxygen (O). At high altitudes ($> 80km$), the lifetime of atomic oxygen exceeds a day, and the energy may be stored as chemical energy. This energy is released as thermal energy when the atom recombines. Due to both horizontal and vertical transport, much of the stored chemical energy is released in the high latitude winter. This process, along with adiabatic heating caused by vertical velocities, is believed to be the main explanation of the warm mesopause temperatures observed in winter (cf. Brasseur and Solomon, 1986). Radiative cooling of the stratosphere and mesosphere is mainly due to vibrational relaxation in the infrared 15 μm band of carbon dioxide (CO_2). In addition, the 9.6 μm band of O_3 contribute to the cooling, especially near the stratospause. Water vapor (H_2O) also contribute, but to a minor extent compared to CO_2 and O_3 (cf. Kuhn and London, 1969; London, 1980; Fels and Schwarzkopf,1981).

The radiative equilibrium described above is not enough to describe the complete thermal budget of the mesosphere. Measurements of atmospheric atomic oxygen number densities, crucial to the understanding of the thermal budget, have to be intensified in order to get the complete picture (Rees, 1989). Other natural minor atmospheric constituents may contribute as well to the coupling between chemistry and radiative transfer. In addition, the general circulation with oscillations due to tidal and planetary waves have to be included. Temporal variations due to gravity waves propagating through the mesosphere is also of importance (Viereck and Deehr, 1989; Viereck, 1991; Hamilton, 1996). External perturbations like variation in the solar flux, energetic particle precipitation, volcanic and anthropogenic emissions are additional sources of concern to the thermal structure of the mesosphere. The increase in greenhouse gases like carbon dioxide and methane in the lower atmosphere are expected to increase the water content and decrease the temperature in the upper mesosphere (Thomas, 1996). Recent model calcu-

Correspondence to: F. Sigernes

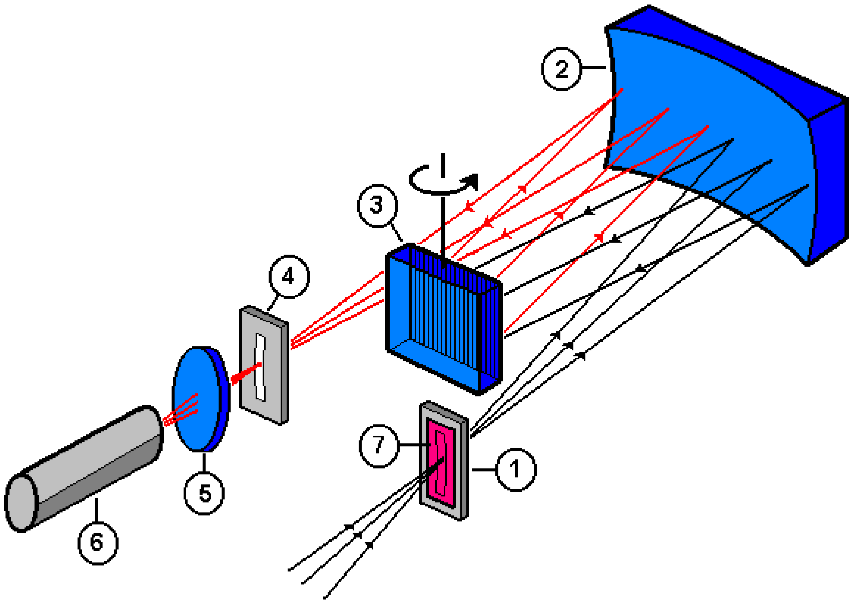

Fig. 1. Three-dimensional optical diagram of the spectrometer. (1) is entrance slit, (2) concave mirror, (3) grating, (4) exit slit, (5) collector lens, (6) photo multiplier tube, and (7) broadband order sorting filter.

lations predict that a doubling of the CO_2 content is indeed necessary in order to explain the long term negative temperature trend of the mesosphere (Akmaev and Fomichev, 1998).

The first indirect indications on a negative temperature trend in the mesosphere was reported by Gadsden (1990). From the increasing number of observed noctilucent clouds (NLC), the average temperature of the summer mesosphere was estimated to decrease at a rate of 2.5 °K per decade. Since ice crystals are believed to be the major part of the NLC particles, a decrease in temperature may be the reason why they are more frequently formed and sighted at lower latitudes than earlier. The fact that the $F2$ peak in electron density in the ionosphere has decreased in altitude, as shown by 39 years of ionosonde observations (Ulich and Turunen, 1997), is consistent with a cooling of the lower parts of the atmosphere. Most of the middle atmosphere is evidently cooling, with values which vary from a few °K at 30-40 km, 10 °K at 50 km and 20 °K at 60-70 km since the mid–60's (Golitsyn et al., 1996). These results where based on direct in–situ rocket measurements. Golitsyn et al. (1996) also reported a record high mesospheric cooling of 30 °K since 1957 using temperature estimates from measurements of the hydroxyl (OH) airglow layer (centered at about 87 km) at Zvenigorod (55.7 °N) and Abastumani (41.8 °N). These last measurements have not been confirmed by results from other stations. Direct observations by the falling sphere technique (Schmidlin, 1991; L$\ddot{u}$bken, 1999), which give temperatures up to approximately 95 km, does not give any indications of such a large decrease of the mesopause temperature (L$\ddot{u}$bken, 2000). Clearly, there is need for more long term temperature measurements from different locations and by different methods.

In addition to the possibility that the above trends signal climatic changes, measurements of parameters such as the temperature and its long and short term variations are important tools in our effort to understand the processes in the mesosphere and thermosphere A change in the temperature will affect the propagation and dissipation of waves, which have a profound influence on these parts of the atmosphere (Bittner et al., 2000). Long term observation series such as those reported by Golitsyn et al. (1996), Bittner et al. (2000) and Duck et al. (2000) are therefore important. As a contribution to our understanding on the temperature in the mesosphere and its variations, we will in this paper report mesospheric winter temperatures derived through spectral measurements of the OH airglow since 1980 from the Auroral Station in Adventdalen (78°N, 15°E).

2 Observational technique

The traditional way and most reliable form of monitoring long term temperatures of the mesosphere is through ground based spectroscopic measurements of the OH airglow layer (Sivjee, 1992; Greet et al., 1998). The molecular band emission of OH was first identified by Meinel (1950). The OH emissions were found to dominate the airglow spectrum from approximately 5200 Å to 4 μm. The OH–airglow layer is, according to in–situ measurements made with rockets, situated around 87km with a thickness of 8 km regardless of season or latitude (Baker and Stair, 1988). The principal source of the vibration excited OH radicals, the ozone-hydrogen mechanism, was suggested independently by Bates and Nicolet (1950) and Herzberg (1951)

$$H + O_3 \longrightarrow OH(v' \leq 9) + O_2 + 3.3eV. \quad (1)$$

The exothermicity is sufficient to excite the OH molecules up to a vibrational level of $v' = 9$. The evidence in favor of the ozone–hydrogen mechanism was obtained in the laboratory by McKinley et al. (1955). Note that both H and O_3 are minor constituents in the mesosphere and lower thermosphere. Their production and loss needs to be understood in order to appreciate the importance of the OH emissions (cf. Rees, 1989). The rotational temperature is inferred from the relative photon emission rates of various lines in a band by constructing synthetic spectra that best match the observations. The obtained rotational temperature is interpreted as kinetic under the assumption that the excited OH molecules are in thermal equilibrium with the atmosphere.

The instrument used in this study, the Ebert-Fastie spectrometer, was constructed by W. G. Fastie at John Hopkins University, Maryland at the beginning of the 70's. Fastie (1952a,b) improved the original design of the monochromator made by Herman Ebert in 1889. In 1978, a $1m$ and a $1/2m$ Ebert–Fastie spectrometer were transferred to the Auroral Station in Adventdalen, Svalbard, from the Geophysical Institute, University of Alaska. Since then the photon counting and computer electronics have been continuously upgraded in order to enhance both the control and performance of the instruments (Sigernes, 1996). Data from these instruments are widely published and recognized. Throughout the 80's, a series of articles have been published on OH temperature measurements from the station (Myrabø at al., 1983; Myrabø, 1984, 1986; Myrabø and Harang, 1988; Sivjee and Hamwey, 1987. The work was mainly on tidal variations, the effect of stratospheric warmings and the seasonal variation. In the late 80's, focus turned onto the effect of gravity waves interacting with the OH layer (Myrabø et al., 1987; Viereck, 1988; Viereck and Deehr, 1989).

Fig. 1 shows the experimental setup of the $1/2m$ focal length Ebert–Fastie spectrometer used for OH measurement in the early 80's. The instrument was later replaced permanently in 1983 by another $1m$ focal length instrument in order to gain spectral resolution. The principal components of the instruments are one large 1m focal length spherical mirror, one plane reflective diffraction grating and a pair of curved slits. The recorded radiance from the sky, is limited by the etendue; the product of the area of the entrance slit and the solid angle field of view. Because of the low intensity of the source, the etendue is made as large as possible. The image of the entrance slit is reflected by one part of the spherical mirror onto the grating. The second part of the mirror focuses the diffracted light from the grating onto the exit slit. When the grating turns, the image of the entrance slit is scanned across the exit slit. A collector lens transfers the output of the exit slit to the front of a photo multiplier tube. The tube is mounted in a Peltier cooled housing. Signals from the tube are amplified and discriminated before sent to the computers counter card. The instrument is setup to scan through the first order wavelength range 8312 - 8745 Å, which corresponds to the $OH(6-2)$ band of the airglow. The band pass or spectral half width of the instrument is 4.8 Å. A red cut-off filter is used in front of the entrance slit to prevent overlapping spectral orders. The spectrometer is fixed in geographical zenith with a field of view of approximately 5°.

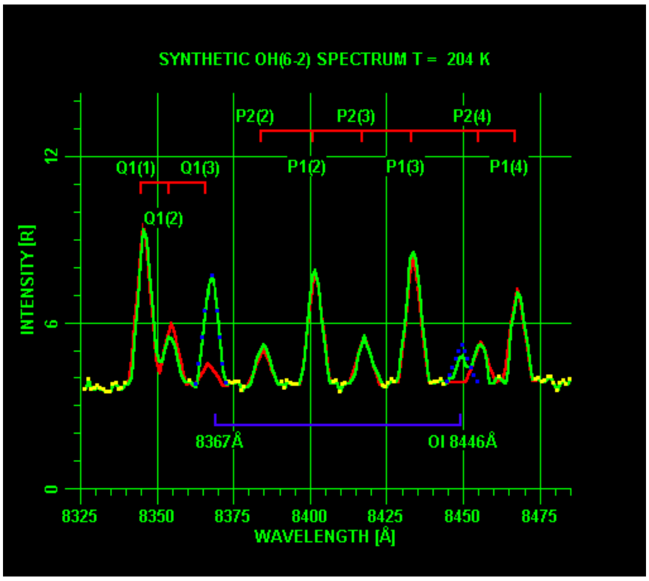

Fig. 2. The measured and synthetic spectra of the OH(6-2) band. Each line is marked and identified according to quantum state. The curve plotted with line color red is the synthetic, green the measured and yellow is the detected background spectra, respectively. The spectral lines plotted with blue color represents emissions caused by aurora.

In order to derive the rotational temperatures of the measured $OH(6-2)$ emissions, it is necessary to produce the corresponding synthetic spectra. The basis for calculating these spectra is given by Herzberg (1950) with term values from Krassovsky (1962) and Einstein coefficients from Turnbull and Lowe (1989). The temperatures can then be derived since the upper populated energy states of the $OH(6-2)$ band are distributed according to the Stefan–Boltzmann distribution. Recent work by Conner (1993) explains in detail how the parameters necessary for generating the synthetic spectra as a function of the rotational temperature can be calculated. The temperature is found by choosing the optimal fit between the measured and the synthetic spectrum through iteration, to minimize the least square error. A computer program (Sigernes and Nielsen, 2000) has been created to implement the above procedure. The program is capable of analyzing large sets of data and is together with samples of raw data freely available to the scientific community.

Fig. 2 shows a typical result of a synthetic fit to a measured spectrum from November 17, 1999. In this case, a temperature of 204°K with a fit value of 0.95 was found. The fit value is defined as

$$Fit = \frac{100 - l}{100}, \tag{2}$$

where l is the least mean square difference between the measured and the synthetic spectrum. This is the same fit function that was used by Myrabø et al. (1983), Viereck (1988) and Viereck and Deehr (1989). A $Fit = 0.8$ represents a relative uncertainty of approximately $\pm 2^\circ K$. In addition, the covariance between synthetic and measured spectrum has been introduced to select time periods where the measurements are not affected by moon light, clouds or aurora. The main problems are moonlight and clouds. The moon does not rise very high above the horizon at Svalbard. However when the sky is cloudy, moonlight is scattered into the spectrometer. As a consequence, the background intensity level rises, which makes the fit of the synthetic spectra poor. A covariance level of 0.9 turned out to work successfully.

3 Results and discussion

The OH daily mean rotational winter temperature from 1980 to 2001 is plotted in figure 3. The data set consists of 624 days and it is not uniformly sampled in time. Each daily mean is calculated as the average temperature of hourly deduced temperatures. A minimum of 3 hrs is used to obtained a reliable daily mean. The selected number of hourly obtained temperatures per day depends mainly on sky conditions as mentioned above. Furthermore, it also depends on whether the instrument was operative or used in another configuration / setup. These are the main factors for the large data gaps seen in the fig. 3. The instrument was not in operation in 81/82 and during the winter season 98/99 the instrument was setup to study proton aurora instead of airglow. In addition, data from 91/92, 92/93 and 96/97 have not yet been retrieved due to problems with storage media and outsourced tape stations. These temperatures will in the near future be restored by using data from other instruments at the station. During the seasons of 80/81 to 82/83 the temperatures were obtained using the $OH(8-3)$ band. Thus, on average, the 80/81 and 82/83 temperatures have been lowered by 5-6°K to make them compatible with the $OH(6-2)$ data from the 83/84 to the 89/90 seasons. Furthermore, the temperatures up to the 89/90 season are calculated using the transition probabilities from Mies (1974). These temperatures are approximately 5 °K too low compared to using the transition probabilities of Turnbull and Lowe (1989). Both these two effects have been accounted for to make the whole series compatible. We have not yet considered all of the corrections. This will be applied to the older data in the future. These possible corrections will not change the conlusion of this particular work.

The average daily winter temperature for 1980 to 2001 is 208°K. The maximum temperature is 257°K in January 14 1983. Minimum temperatures down to 150 °K are seen in January of 1983 and 1991. The standard deviation is close to 16°K , indicating that the mesospheric temperature variations over Svalbard in winter are extremely high. The day to day temperatures show large maxima lasting for a few days to a week or longer. Note that large, short time-scale variations in the temperature may occur within each daily average. Myrabø et al. (1983) detected large amplitude variations of both the OH intensity and the temperature, and were the first to suggest that these variations are actual variations rather than noise. Amplitudes of 20 °K in the internal gravity wave period range (minutes to hours) with extremes up to 70 °K within a few hours are usual.

On a larger time scale, the daily mean temperatures show a wave–like pattern with deviations from the mean, each lasting for about 3–10 days. Myrabø et al. (1986) analyzed the seasonal variation from 80/81, 82/83, 83/84 and 84/85 and reported a semidiurnal tide with an amplitude of 3 °K in the December 1982 data. By early January 1983 the tide increased in amplitude to 10 °K. The increase was correlated with a stratospheric warming (Labitzke, 1981; Labitzke et al., 1985). A cooling of the mesosphere around New Years with a total amplitude of 20–30 °K tends to occur each year. In addition, Walterscheid et al. (1986) reported a semidiurnal tide of 13°K on day 359 in 1984. The tide was called pseudo–tide, believed to be generated by gravity wave mean flow interactions modulated by the solar–driven tide. Nielsen et al. (2001a) reported a cooling of 30 °K in December 1987, which again was associated with a stratospheric warming. The strength of the diurnal and semi diurnal tides were at that time 30 and 15 °K, respectively.

In general the mesosphere has a winter warming, peaking near solstice. It is interspersed with a period of 2–5 days when the temperature drops almost down to the summer minimum. These minimums are probably associated with stratospheric warmings that occur with varying strength all winter. The 80/81 to 84/85 seasons have a relative low temperatures in late December followed by a warm mesospause in early January. There are exceptions, the 85/86, 89/90 and 99/00 seasons show the opposite tendency. There are also more quiet years with smoother seasonal variation. A clear example is the 99/00 season. There is in other words no distinct identifiable shape of the seasonal temperature variation from year to year. Comparison with other OH-measurements to similar measurements performed at mid and low latitude stations show that both short period and seasonal temperature variations are consistently larger in the OH airglow over Svalbard. Similar measurements from Antarctica show smoother temperature variations compared to Svalbard. Both the daily and weekly variations of the temperatures exhibit variations on the scale of 10 °K. Nielsen (2001b) showed that these variations have a statistically significant connection to the variation of the average strength of the stratospheric polar vortex at $10hPa$ height, which corresponds well with the general circulation theory and the theory of gravity wave propagation. Some of these variations could also be caused by mesospheric planetary waves as discussed by Nielsen et al. (2001a). No auroral effects have so far been reported / observed to influence the OH temperature series.

Another thing to consider is the seasonal variation with the direct impact of the solar UV flux on mesosphere. Sahai et al. (1996) reported a correlation of 0.73 between the OH temperature over Calgary (51 °N, 114°W) and the 10.7 cm solar flux using 87/88 and 1990 to represent seasons with low and high solar activity, respectively. The 1990 winter temperatures were approximately 30 °K warmer than the 87/88 temperatures, which is in the order of 10 °K higher than model calculations by Huang and Brasseur (1993). At the moment we do not see any obvious connection in our data. Further studies are indeed needed to investigate the correlation be-

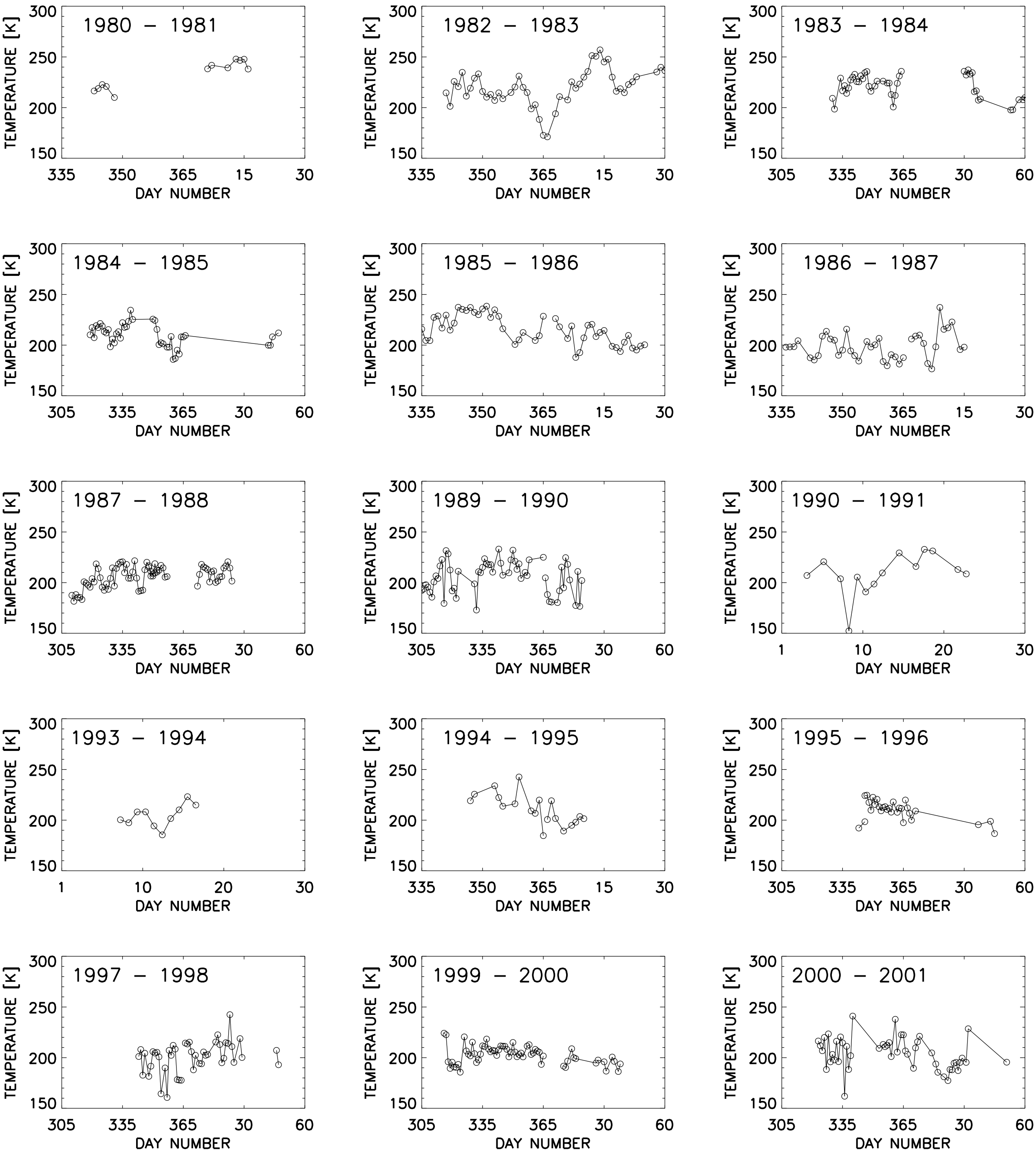

Fig. 3. Daily mean temperatures (marked with circles) as obtained from OH band night airglow emissions for the 80/81, 82/83, 83/84, 84/85, 85/86, 86/87, 87/88, 89/90, 90/91, 93/94, 94/95, 95/96, 97/98, 99/00 and 00/01 winter seasons at the Auroral Station in Adventdalen, Svalbard.

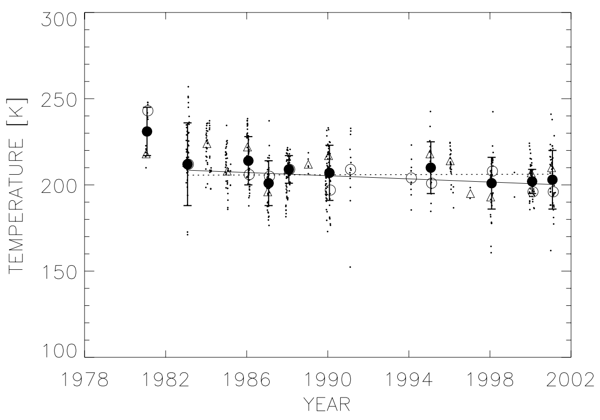

Fig. 4. Seasonal mesospheric OH airglow winter temperature trend over Svalbard 1980–2001. The daily mean temperatures are plotted as small dots. The monthly mean temperatures are plotted for December (triangles) and January (circles). The seasonal averages (filled circles) have the standard deviation plotted as error bars. The solid line represents the linear trend. The dotted line is the estimated linear trend using weights equal to number of days per season divided by the standard deviation.

tween the direct influence of the solar flux and the mesospheric temperatures.

In order to determine the temperature trend with time, we have to correct for variation with solar cycle. As seen above this variation affects the annual variation and is mainly implied by the variation in strength of the stratospheric warmings with solar cycle. We know that the stratospheric warming is associated with mesospheric cooling. A set of criteria is necessary in order to separate the stratospheric–cooled mesosphere from the normal yearly variation:

– Take upper envelope of annual daily average temperatures.

– Compare to the 10 mb temperature and exclude periods of stratospheric warmings.

– Use only temperatures from days when the semi–diurnal amplitudes are less than 6°K.

– Compare standard variation of yearly average temperature to the strength of stratospheric warmings.

A first cut would be to average the data for each year independent of whether stratospheric warmings occur or not. Figure 4 shows both the monthly and daily temperatures. The overall seasonal or yearly mean winter temperature is calculated as an average between the monthly mean temperatures of December and January. A linear trend is calculated using the yearly averages, expect for the 80/81 season. This season was excluded since a criteria of minimum 14 days per season is applied. The trend is calculated using a polynomial fit of first degree with weights defined as number of days per season divided by the standard deviation for each the yearly average temperatures. This trend is compatible with zero: +0.03°K/y±0.002°K/y. The uncertainty of the trend estimate is calculated by the bootstrap method (Efron and Tibshirani, 1993). If we do not apply weights, the trend becomes -0.4°K/y±0.09°K/y. Note that the errors in each point of the polynomial fits with and without weights are of the order of ±1°K and ±3°K, respectively. Our trend estimate is so far definitely not comparable to the large values up to -1 °K/y at mid latitudes reported by Golitsyn et al (1996). The nearly zero temperature trend in the polar summer mesosphere reported by L*ü*bken (2000) seems to be more compatible with our results. We conclude that there is no obvious temperature trend in the polar winter mesosphere over Svalbard during the last 20 years.

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