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Temperatures using radar-meteor decay times.

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Abstract. Experimental studies of the temperature and pressure dependence of the ambipolar diffusion coefficient in the mesopause region have been undertaken by studying meteor trail decay times with radars at a variety of sites in North America, with latitudes between 75N and 35N. The site at Resolute Bay, Canada, has proven especially useful, due to the wide range of mesopause temperatures experienced at that site between summer and winter. Theoretical predictions have been confirmed, and an algorithm is outlined which permits meteor decay times to be used to determine absolute measurements of mesospheric temperatures.

1. Introduction

As meteoroids enter the Earth's atmosphere, they form approximately cylindrical plasma trails with considerable electron content. These trails are capable of reflecting radio waves which impinge upon them, and so a suitably configured radar can be used to detect and interrogate them. The trail forms quickly and then expands radially as time progresses. Assuming that ambipolar diffusion is the main trail expansion mechanism in the early stages of growth, then the backscattered radar signal for an underdense meteor appears as a sudden leap in amplitude to an initial value A_0, followed by an amplitude decay according to

$$A(t) = A_0 e^{-(16\pi^2 D_a t)/\lambda^2} = A_0 e^{-\ell n2 \frac{t}{\tau_{1/2}}} \qquad (1)$$

where t is time, λ is the radar wavelength, D_a is the "ambipolar diffusion coefficient", and $\tau_{1/2}$ is the time for the amplitude to fall to one half of its maximum value. $A(t)$ is the received field strength at time t, with t=0 being the time at which the meteor signal first appears. The value $\tau_{1/2}$ is typically in the range 0.01 to 0.5 seconds for a radar operating at a frequency in the range 30 to 50 MHz. (e.g. see *Hocking et al.*, 1997, and references there-in). This paper deals exclusively with such underdense meteors.

By measuring the half-amplitude decay time $\tau_{1/2}$ of the meteor signal, it is possible to estimate the parameter $D_a = \lambda^2 \ell n2/(16\pi^2\tau_{1/2})$. This parameter in turn depends on the atmospheric temperature and pressure, and *Jones and Jones* [1990] have predicted that D_a can be determined as

$$D_a = K_{amb}\frac{T^2}{P}. \qquad (2)$$

Specific details about K_{amb} have been outlined in *Hocking et al.*, [1997] and *Chilson et al.*, [1996]. *Hocking et al.* have demonstrated application of this theory to determination of the parameter $\chi^\circ = T/\sqrt{P}$, and shown that experimental measurements of this parameter using meteor decay times agree reasonably well with CIRA (Cospar International Reference Atmosphere) estimates of χ°.

However, despite the good progress in determination of D_a and thence χ°, as demonstrated by *Hocking et al.* [1997], the methods discussed there do not produce an absolute measure of temperature. It is possible to convert χ° to a temperature if model pressures are used, like those specified by *Fleming et al.*, [1988] (hereafter *F88*), but these can at times have insufficient accuracy or reliability. We therefore seek a technique which allows absolute determination of temperatures, independent of knowledge of P. In addition, the assumed relation (2) is still a largely theoretical derivation, and needs stronger confirmation.

Two objectives are addressed. We demonstrate using data from a meteor radar installed at Resolute Bay in Northern Canada (75N, 95W) that D_a is indeed consistent with (2), and then we show that this knowledge allows us to develop a strategy for absolute determination of the temperature at the height of maximum meteor count rates (typically 86-90 km, depending on radar frequency, season and location).

2. Instrumentation

We utilize data from three separate meteor radars, operating at four separate sites distributed throughout North America in the years 1997-1999. The first was the CLOVAR instrument situated at London, Ontario, which has been described by *Hocking et al.*, [1997]. Its frequency of operation was 40.68 MHz. The second was a similar radar to the CLOVAR instrument, but it used 5 receivers, one for each receiving antenna (in contrast to CLOVAR, which multiplexed 5 antenna signals through one receiver). Called SKiYMET (allSKy interferometric METeor radar), this second radar operated at a frequency of 35.24 MHz and was installed at London, Ontario, from November 1997 to February 1998, then at Saskatoon (52° N, 117° W) from April to July 1998, and finally at Albuquerque, N.M. (35° N, 107° W) from September 1998 to the present. The third radar was a system similar to that described by *Hocking and Thayaparan* [1997], but sited at Resolute Bay. It differed from the other two radars primarily in receiver antenna layout and frequency. For reception it utilized only four receiving antennas, with signals being multiplexed through a single receiver. Although inferior to the improved 5-antenna interferometer used with the SKiYMET systems, it was still quite adequate for the studies described here-in. In the case of the newer 5-antenna interferometer, angular accuracies of ± 1-2° in meteor angular location were possible, whilst the 4-antenna system has a resolution of typically ± 2-3°.

In all experiments, our objectives were the same: to locate as many meteors as possible, calculate their angular position in the sky, use their known range for determining their height, and then use correlation techniques to determine their decay times. Typically the Resolute Bay radar detected 150-600 meteors per day, the CLOVAR radar typically 600-1200 per day, and the SKiYMET radar typically 1000-2500 per day, depending on background noise level and general meteor activity. Using these data, scatter plots of height (z) vs $log_{10}(1/\tau_{1/2})$ were prepared, as demonstrated in Fig. 1. These particular graphs were produced using data from the Resolute Bay radar, for typical winter and summer conditions. Similar graphs for the CLOVAR radar were also shown in *Hocking et al.*, [1997]. The subsequent analysis will now be described.

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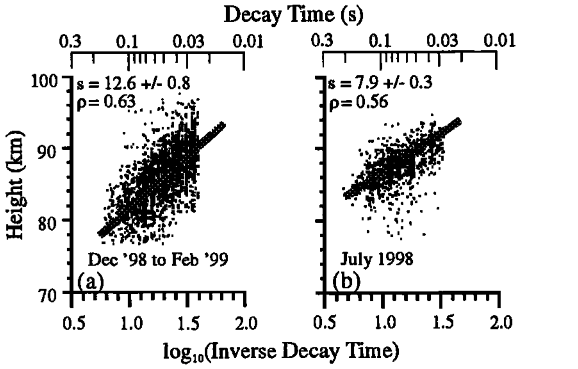

Figure 1. Scatter plots of meteor heights vs. logs of inverse decay times for Resolute Bay for (a) Winter 1998-9 and (b) July 1998. The slope of the best fit line is s, and the correlation coefficient is ρ.

3. Test of the dependence of D_a on Temperature and Pressure

Our first agenda here will be to perform a test of equation (2), and then use the results of that test to develop a new method for determination of absolute temperatures.

To begin, it is clear from figs. (1a) and (1b) that the slopes of the best fit straight line for these two scatter-plots are very different; indeed they differ by almost a factor of 2. The errors in the slopes are of the order of 4-6%. We will demonstrate that these different slopes relate to the mean temperature at the height of maximum meteor activity. At many sites, the differences in slope between winter and summer can be slight, but because high northern latitudes exhibit a large temperature range (from 210 K to 130 K) between seasons [e.g. see *Lübken and von Zahn*, 1991 ; hereafter LV91] the difference is very substantial. We utilize this feature to demonstrate that D_a is proportional T^2/P to within experimental error. To do this we presume that $D_a \propto \frac{T^\Omega}{P^\iota}$. Then

$$log_{10} D_a = \Omega \; log_{10} T - \iota \; log_{10} P + c_1, \tag{3}$$

where c_1 is a constant. If one moves downward over a height interval of about one scale height (about 7km), P changes by typically a factor of 2.7, whereas even in extreme circumstances T will change by at most a few percent. Hence we recognize that P is a rapidly changing function of height relative to T, so we may write

$$log_{10} D_a = -\iota \; log_{10} P + c_2(z, T) \tag{4}$$

where c_2 is a relatively weakly varying function of height and temperature relative to the pressure term. For an isothermal atmosphere, $P = P_0 e^{-\frac{mg}{kT}z}$, where g is the acceleration due to gravity (9.49 ms^{-2} at 90 km altitude), m is the mass of a "typical" atmospheric molecule, k is Boltzmann's constant and T is the temperature. If we utilize this, and also recognize that D_a is proportional to $1/\tau_{1/2}$ (equation (1)), we may then write

$$z = \frac{1}{\iota} \frac{1}{log_{10} e} \frac{kT}{mg} \log_{10} \left(\frac{1}{\tau_{1/2}}\right) + c_3'(T, z) \tag{5}$$

where c_3' is a weakly varying function of T and z. We will write this as

$$z = S_m \log_{10} \left(\frac{1}{\tau_{1/2}}\right) + c_3'(T, z). \tag{6}$$

We therefore see that, to a reasonable approximation,

$$T \simeq \iota \frac{log_{10} e \; mg}{k} S_m \simeq 14.3 \iota S_m \tag{7}$$

It should also be recognized that the best-fit lines in fig. 1 are simple fits to the raw data. However, *Hocking et al.*, [1997] demonstrated that this is not the best estimate of the true mean profile of height vs. $log_{10}(1/\tau_{1/2})$, due to natural biases which occur in the detection process. In section 3.3 of that paper, a procedure was outlined which applies a deconvolution bias adjustment. When this procedure is applied, it steepens the slope by typically 10% to 20%. We have applied a least-squares fit to the raw data, without bias adjustment, in fig. 1, solely for display purposes. However, in all future discussions we utilize a best-fit line which has this adjustment incorporated. It is this bias-adjusted slope which we consider to be S_m in equation (7).

Fig. 2 shows a scatter plot of T vs S_m for 11 different months using the Resolute Bay data (10 months from 1998, and one from July 1997). We have used the values of temperature as determined by *LV91* at 69°N for those months for which those authors had data, and the values from the model of *F88* for those months in which *LV91* did not make measurements. The months of March and November in 1998 were not used due to unavailability of meteor data. We have then fitted a further least-squares line to fig. 2.

The best-fit slope for the data shown in fig. 2 was 14.09 $\pm$ 1.7, which corresponds, after comparison with (7), to a value for ι of 0.99 $\pm$ 0.12. Thus this is consistent with the assumption that $\iota = 1$. We therefore consider that these data demonstrate experimentally that D_a is inversely proportional to pressure, as predicted by (2): we shall henceforth take $\iota = 1$.

Our next step is to determine the value of Ω. We return now to equation (3), but assume $\iota = 1$. Then we see that if $log_{10}(T)$ is plotted as a function of $log_{10}(D_a P)$, we expect a slope of $\frac{1}{\Omega}$. We can determine D_a experimentally by determining $\tau_{1/2}$ and thence D_a through equation (1), and we can use the same temperatures as those used in fig. 2. However, we need absolute measurements of the pressure P, which represents an area of uncertainty.

In order to ascertain a suitable pressure model, we have examined pressure data from other sources as a function of month. We have used two sources: firstly the empirical model of *F88*, and secondly the experimental data of *LV91*. Using 88 km altitude as an example, the Fleming model shows a summer maximum of about 0.35 to 0.4 Pascals, and a winter minimum of about 0.2 Pa. In contrast, *LV91* shows a generally constant value of pressure throughout all seasons, with all values being confined between 0.2 and 0.25 Pa. These are therefore very different in behaviour, and we need to decide which is best to use for Resolute Bay. When

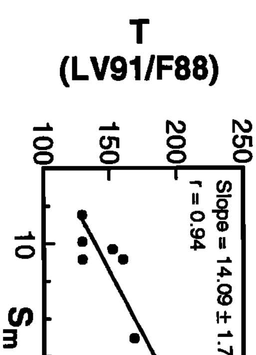

Figure 2. Bias-adjusted slopes of the best-fit lines to scatter-plots like those shown in fig. 1, relative to known 88 km temperatures, for 1998 at Resolute Bay.

we plot T vs $(D_a.P)$, we get the two "scatter plots" shown in Fig. 3. Remarkably, the *F88* profiles trace out an ellipse as a function of season. This is very significant - it suggests that T and $(D_a.P)$ show a generally sinusoidal variation as a function of season, but are out of phase by 90°. This is unreasonable, so we cannot use these pressures. In contrast, the data using *LV91*'s pressures (fig. 3b) show a straight line trend, with slope 0.56 ± 0.07. The 95% confidence interval limits are 0.41 and 0.71. This corresponds to a value of Ω of 1.8 ± 0.2, with 95% confidence of being in the range 1.4 to 2.4. We therefore surmise that (i) the *F88* (and CIRA) pressures at Resolute Bay are a poor representation of the true seasonal cycle, (ii) *LV91*'s values of pressure are reasonable, and (iii) Ω is consistent with a value of 2 within expected error. We therefore claim that our results support the original modeling results of *Jones and Jones* [1990], and we accept that $D_a \propto T^2/P$.

4. Absolute Temperatures

We now use these results to determine T. However, extra care is needed if we require good accuracy, so we no longer assume that the atmosphere is isothermal. Rather, we let $T = T_0(1+\alpha z')$, where the quantity α relates to the mean temperature gradient. We should emphasize that ignoring this term altogether produces errors in the absolute temperature of less than 5-10%, but we can improve our estimates of T if we include it. We therefore now examine the expected temperature gradients as a function of season and latitude.

During much of the year, especially the non-summer months, the mesopause at mid to high latitudes is above the region of peak meteor activity, so that the temperature profile through that region is moderately linear, with a fairly well defined gradient, at least when averaged over periods of days. In summer, occasions exist when the mesopause dips below the meteor peak height, so that $\frac{dT}{dz}$ becomes positive. By studying these gradients using the model of *F88*, and also from experimental rocket and lidar studies [*LV91*, *States and Gardner*, 1999], we have found that the following expression gives a generally reasonable estimate of the gradient as a function of latitude and time at the height of peak meteor count rates;

$$\left[\frac{dT}{dz}\right]_{av} = -1.5 - \left[-2.5\, exp\left\{-(\theta-45)^2/200\right\} \right.$$
$$\left. +1.5\, exp\left\{-(\theta-90)^2/1350\right\}\right] \times exp\{-\#^2/3200\} \qquad (8)$$

where θ is the latitude in degrees, and # is the temporal displacement in number of days from mid-June in the northern hemisphere (or displacement in days from mid December in the southern hemisphere). This embodies a mean temperature gradient of about -1.5 K/km in winter (and the equinoxial months closest to winter) at all latitudes, a gradient of about -1.5 K/km in the equatorial regions in summer, a tendency to zero or positive gradients at mid-latitudes in summer (low mesopause height), and a tendency to very steep negative gradients in the polar regions in summer.

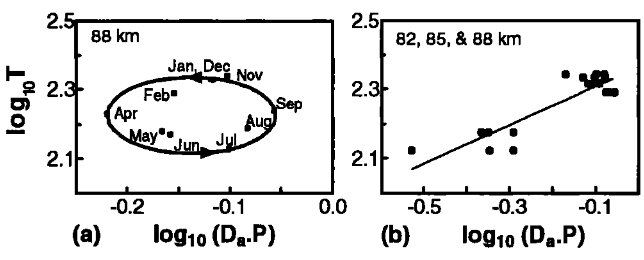

Figure 3. Log-log graphs of $(D_a \cdot P)$ vs. temperature for Resolute Bay, using two different estimates of pressure, for the heights indicated. In (a), the pressures of *F88* were used whereas in (b), the pressures presented in *LV91* were used.

Having developed an expression for the gradient, we now utilize it in the following way. Assuming for generality that $D_a = K_\Omega T^\Omega/P$, and defining a vertical coordinate z' which is zero at the height of peak meteor activity, and recognizing that generally $P = P_0 exp\{-\int_0^{z'} \frac{mg}{kT} dz''\}$, we may write

$$log_{10}(D_a) = \Omega log_{10}\left[T_0(1+\alpha z')\right]$$
$$+log_{10}e\frac{mg}{k}\int_0^{z'}\frac{1}{[T_0(1+\alpha z'')]}dz'' + \text{constant.} \qquad (9)$$

where $\alpha = \frac{1}{T_0}\frac{dT}{dz}$. If we now differentiate this equation with respect to height, and evaluate it at $z' = 0$, we have

$$\frac{1}{S_m} = log_{10}e\;\Omega\;\alpha + log_{10}e\frac{mg}{k}\frac{1}{T_0}. \qquad (10)$$

S_m is the slope of the graph of z' versus $log_{10}D_a$, or equivalently the slope of the graph of z' versus $log_{10}(1/\tau_{1/2})$, after bias adjustment. Solution of (10) with Ω =2, and α determined from (8), allows determination of T_0.

5. Results

We have determined S_m and thence T_0 from (10) for all of our radars, and compared our results to other experimental data measured at similar sites. For example, our measurements at London, Ontario, have been compared to spectrometer and lidar measurements of *She and Lowe*, [1998], as well as lidar measurements from Urbana [*Senft et al.*, 1994]. Our measurements at Resolute Bay have been compared to those of *LV91* at Andoya (69 N). Our measurements at Saskatoon and Albuquerque have been compared to *F88*, after addition of about 5K to the *F88* data. This addition was included to compensate for a bias in *F88* and CIRA temperatures which seems to occur at mid-latitudes, as evidenced by *She and Lowe*, [1998], and *Senft et al.*, [1994]. A graph of our monthly data vs. these other measurements is shown in fig. 4a. Clearly the data are closely aligned along the line of zero offset and unit slope. However, there is also a tendency for the meteor radar to underestimate the "true" value at low temperatures, and overestimate (very slightly) at higher ones. A best-fit line shows that

$$T_{true} = 0.774\; T_{meteor} + 42.8. \qquad (11)$$

The reason for this slightly skewed line is not clear. If ι were not exactly unity, or Ω was not exactly 2.0, this could alter the fits. This is possible, since the errors on these exponents deduced earlier were of the order of 10 to 20%. We cannot fine-tune our procedures to any higher degree, so we recognize equation (11) as a "calibration curve". The fit has a correlation coefficient of 0.9 and a typical vertical scatter about the line of $\pm 7K$. At least some of this scatter is due to the fact that the reference temperatures are from different sites and different years, so the intrinsic error in our data must be considerably less than $\pm 7K$.

Therefore, as a final step, we apply equation (11) to our meteor temperatures to produce our final estimates of T. Given that the typical vertical spread of points about this best-fit line is $\pm 7K$ (only part of which is due to our technique), it is evident that it is possible to make measurements of monthly mean temperatures using this procedure with precision of the order of 4 to 8 K.

As justification for this statement, we present fig. 4b, which shows monthly mean temperatures measured at London, Ontario, as averaged over the period March 1997 to February 1999 inclusive. Superposed are temperatures for

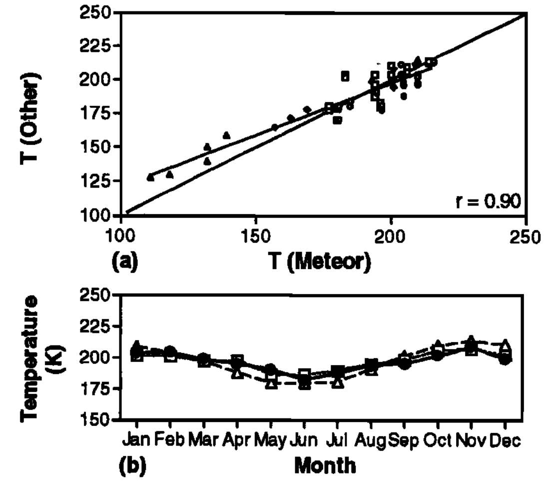

Figure 4. (a) Scatter plot of the "temperature" deduced from equation (10) (abscissa) compared to temperatures at similar sites from other techniques like spectrometers, lidar and empirical models (ordinate), for ~88 km altitude. The solid triangles refer to Resolute Bay (75N, 1998-9) compared to LV91 *(69N)*, the diamonds to Saskatoon (April - July 1998) compared to *F88*, and the inverted triangles refer to Albuquerque, NM (Dec 1998 - Feb 1999) compared to *F88* for 35N. The CLOVAR data (London, Ont: 43N) were compared to OH data from London for 1993 and lidar data from Fort Collins (41N) for 1993 [*She and Lowe, 1998*], and lidar data from Urbana (40N) for 1992-3 [*Senft et al., 1994*]. The open circles use CLOVAR data from 1997, and the open squares are for 1998. For some months comparisons were not possible for certain instruments, due to unavailability of data; the graph contains 75 points in total.
(b) Monthly mean temperatures at ~88 km over London, Ont. (43N), for 1997-8, after application of the calibration equation (11) (solid circles). For reference the lidar temperatures at 40N for 1992-3 [triangles: *Senft et al., 1994*] and 1996-7 [squares: *States and Gardner, 1999*] are also shown.

Urbana (40N) as a comparison. These data were produced during 1992-3 [*Senft et al.*, 1994] and 1996-8 [*State and Gardner*, 1999]. *She and Lowe*, [1998] showed similar values. It should be emphasized that the data due to *State and Gardner*, [1999] were not used in the original calibration, so to some extent this comparison is independent of the earlier calibration. Absolute values are clearly very similar. Differences between lidar and meteor temperatures are no worse than differences between lidar measurements which used different sampling strategies in different years.

6. Discussion

Previous ground-based temperature determinations have often used optical methods (lidar, spectrometers), and these can have better precision than the accuracies cited above. However, they suffer because (i) they can generally only obtain useful data at night, during cloud-free and low moonlight conditions, and (ii) in the case of passive optical techniques like spectrometers, the actual height of the emissions can be unknown to within 5-10 km altitude. Daytime optical instruments are rare and expensive [*e.g. States and Gardner*, 1999]. In the case of the meteor technique, the mean height at which the temperature is determined is well known, and measurements are possible at all times, including daytime and during cloudy periods. Given that tidal variations can have amplitudes as high as 15K, and are often locked to local time, this can mean that the "daily average" temperatures determined by optical methods can have systematic errors as high as 10 K [e.g. see *States and Gardner*, 1999]. Thus we consider that the methods outlined here-in can serve a useful role as atmospheric temperature monitors in the future. The 24-hour coverage available, and the relatively low cost of meteor systems, are distinct advantages.

7. Conclusion

It has been illustrated that it is possible to use underdense meteor decay times collectively to make reasonable estimates of the temperature at the height of maximum meteor detectability, which is in the range 86 to 92 km (depending on frequency) and is easily measured. The typical accuracy of these measurements is of the order of 4 to 10 K, depending on circumstances. The method is more accurate in the non-summer months, when the temperature gradient can be better and more reliably represented. A strong advantage of the method lies in its ability to make daytime measurements.

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