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Use of MST radars to probe the mesoscale structure of the tropopause

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ABSTRACT

Profiles of the power scattered at vertical incidence by an MST radar have been compared with temperature profiles measured by radiosondes launched nearby. The results are consistent with a standard scattering model for low and medium scattered power, but deviate from the model at high power. Tropopause heights have been derived from the radar data by two methods: the maximum power and maximum power gradient. The two criteria appear to be equally effective as predictors of the tropopause, despite the theoretical expectation that the latter should be superior. Radiosonde tropopauses which were highly indefinite (no marked change in lapse rate between troposphere and stratosphere) were associated with a shallower minimum than normal in the radar power profile in the upper troposphere. This ability to detect highly indefinite tropopauses shows that an MST radar may be used to monitor continuously the structure of the thermal tropopause.

1. Introduction

The possibility of measuring the height of the tropopause by VHF radar, at frequencies around 50 MHz, was introduced by Gage and Green (1979). They noted that the radar echoes from a vertically-pointing beam are strongly enhanced in the lower stratosphere. These echoes are specular in nature: the echo power decreases rapidly with angle as the beam is pointed away from the zenith. A typical echo power profile from a vertically-pointing beam is presented in Fig. 1. This clearly shows a minimum in the upper troposphere and a secondary maximum just above the tropopause (here at 8.5 km), with a decrease in power with height in the stratosphere.

A theoretical model for the enhanced VHF echoes at vertical incidence was presented by Gage and Balsley (1980) and Gage et al. (1981). Termed Fresnel scatter, the model relied on horizontal coherence of temperature irregularities in stable regions of the atmosphere, i.e., a consistent pattern of vertical variations in temperature over a horizontal region commensurate with a Fresnel zone of the radar, $\sqrt{Z\lambda/2}$, where Z is the range and λ is the wavelength of the radar. Such irregularities were postulated to arise from anisotropic turbulence and to persist after the turbulence dissipated, and have since been observed directly by aircraft transects of the tropopause region (Salathe and Smith, 1992). Vertically-reflected radar power is reinforced by the horizontal coherence, consistent with the marked specularity of the observed echoes.

The radar reflectivity is proportional to the Fourier component of the vertical gradient in refractive index, n, at half the radar wavelength: $\partial n/\partial z\,|_{\lambda/2}$. By appealing to the universal spectra commonly associated with turbulence, Gage et al. postulated that the Fourier component of interest would be proportional to the average value of $\partial n/\partial z$, or $\bar{M}$, sometimes termed the potential refractive index. Thus, the power received $P \propto \bar{M}^2$:

$$P = \frac{\alpha^2 P_t A^2}{4\lambda^2 Z^2} \frac{(\Delta Z)^2}{16} F^2(\lambda)\, \bar{M}^2, \qquad (1)$$

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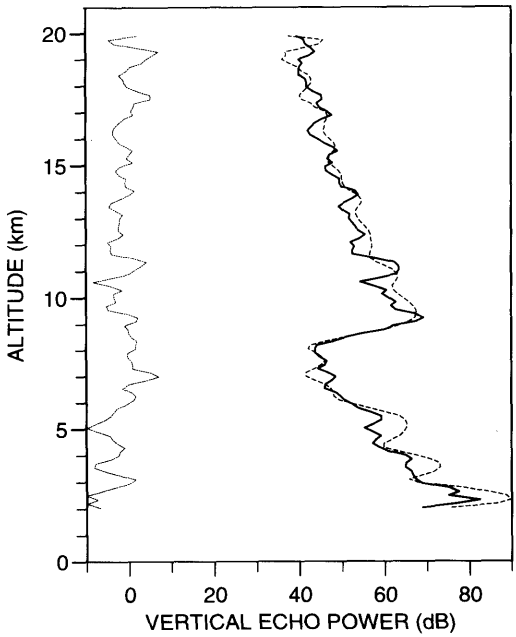

Fig. 1. Examples of measured (solid line) and model (dashed line) echo power profiles from the vertically-pointing beam of the Aberystwyth MST radar. The model values were calculated from linear regression of the data in Fig. 2. Differences between the two profiles are shown by the close dotted line.

where P_t is the transmitted power, A the effective area of the antenna array, ΔZ the range resolution, α an efficiency factor and $F(\lambda)$ an empirical parameter. In the dry air of the upper troposphere and lower stratosphere, humidity fluctuations do not affect the reflectivity at 50 MHz (Doviak and Zrnic, 1984).

$$M = -77.6 \times 10^{-6} \frac{p}{T} \frac{\partial \ln \theta}{\partial Z}, \tag{2}$$

where p is pressure in mb, T the temperature and θ the potential temperature. If all the system constants are combined, and F is considered constant with range,

$$P(Z) = \frac{B}{Z^2} \left(\frac{p}{T} \frac{\partial \ln \theta}{\partial Z} \right)^2, \tag{3}$$

where B is a constant for a particular radar. This expression provides a direct link between the power received and the atmospheric static stability and is the basis of the study reported here.

Gage et al. (1985) further discussed the parameter F, suggesting that it decays exponentially with height. This led to an expression of the form:

$$P(z) \propto \bar{M}^2 \Delta Z \exp\left(-\frac{Z}{H} \right). \tag{4}$$

As shown in Section 2, the present study supports the model of constant F as in (3), rather than exponential decay. Further modifications were introduced by Hocking and Rottger (1983) in cases where M varies rapidly in the vertical.

Because the temperature profile in the upper troposphere is normally near to the adiabatic lapse rate, the static stability $\partial \ln \theta / \partial Z$ is small and the echo power weak. At the tropopause the static stability increases suddenly to values typical of the stratosphere, causing a sudden increase in P. Above this altitude, the temperature in the lower stratosphere (in mid-latitudes) is roughly constant with height. It is then readily shown that $\partial \ln \theta / \partial Z$ is also a constant (equal to $g/C_p T$). $P(Z)$ therefore decreases with height as $(p/Z)^2$, i.e., with roughly half the atmospheric scale height. Consequently, there is a maximum in radar echo power just above the tropopause (Fig. 1).

An obvious application of the observations and theory outlined above is to determine the tropopause height on a continuous basis from radar profiles. To do this it is necessary to verify that the tropopause height derived in such a manner is consistent with that derived from radiosondes, and several studies directed to this end have been reported in the literature. Using a radar with 1 km range resolution, Gage and Green (1979) determined the tropopause height by a subjective criterion and compared their results with radiosonde profiles from a launch site 50 km away. From 30 cases, they derived a correlation coefficient of 0.84 and a r.m.s difference of 0.71 km. Gage et al. (1986), developed an objective criterion and applied this to data from two radars with different range resolutions. Using a resolution of 750 m they found (from 13 radiosonde comparisons) a mean difference of 150 m and a standard deviation of 550 m. For a range resolution of 2.25 km, larger standard deviations were found (700–1000 m).

The purpose of the current work is, firstly, to see whether the tropopause may be located more precisely with a high resolution radar, and secondly to use the radar to derive a parameter for the sharpness of the tropopause. This may be compared to the sharpness of the thermal tropopause as defined by Price and Vaughan (1993). They introduced 3 categories of tropopause sharpness, depending on the height Δh between the WMO tropopause and a lapse rate of 6 K km^{-1}: definite ($\Delta h < 0.5$ km), intermediate (1.2 km > $\Delta h > 0.5$ km) and indefinite ($\Delta h > 1.2$ km). Later, more extensive studies (Jones et al., 1994) refined the lapse rate criterion from 6 K km^{-1} to 5 K km^{-1}, which is used in this work. Indefinite tropopauses were found to correspond to regions of stratosphere-troposphere exchange, and a motivating factor for the work described here was to identify such regions with the (continuously-sounding) MST radar.

2. Comparison of scattering model with measurements

The model described above contains empirical elements, such as assuming that $\partial n/\partial z|_{\lambda/2} \propto \bar{M}$. Its validity was therefore investigated using vertical power profiles from the Aberystwyth MST radar (Slater et al., 1992) together with temperature profiles measured by radiosondes launched from a site 5 km away. These data were used to construct scatter plots of radar versus model power (as derived from the radiosondes). During the period November 1991–April 1992, 32 coincident profiles were obtained. Radar power was measured with 300 m vertical resolution, averaged for 1 hour around the sonde launch time. The potential temperature profiles, available with 30 m resolution, were smoothed using a low-pass Butterworth filter with variable cutoff; the best agreement between model and measurements was obtained with a cut-off vertical wavelength of 750 m (corresponding to an effective vertical resolution of 375 m).

Humidity affects the refractive index in the lower part of the troposphere at the radar wavelength of 46.5 MHz (Doviak and Zrnic, 1984, p. 10), so eq. (3) should not be expected to apply there. The data in this study were therefore restricted to heights above 8 km. Since the radiosonde balloon drifts away from the launch site with time the best agreement should apply at the lower heights and an upper limit of 14 km was used. Profiles of $10 \log_{10}(M/z)^2$ were generated for each ascent, and paired with the corresponding radar echo power (in dB). For all 32 ascents this produced 1251 data points (Fig. 2). A linear regression analysis yielded a slope of 0.80 and a correlation coefficient of 0.81.

However, when the data are examined more closely, a simple linear model appears inappropriate. The data are consistent with a slope of 1 up to about 55 dB, but the slope thereafter decreases. The scattering model thus appears consistent with the data at fairly low echo power (when the degree of specularity is fairly small) but a non-linear relation between power and $\bar{M}^2$ would appear more appropriate at high echo power. Alternatively, enhanced scattering from active layers of turbulence (rather than fossilised structure in M^2) is responsible for the larger values of echo power.

There is considerable scatter in the points plotted in Fig. 2. To estimate this random component, a series of polynomials (up to degree 4) were fitted to the data. The mean-square residual from all polynomials of order 2 or greater was 3.93 dB, while that from the linear regression was 4.0 dB. These residuals are considerably larger than the measurement errors, estimated as < 1 dB for both the radar and model echo power. Most of the

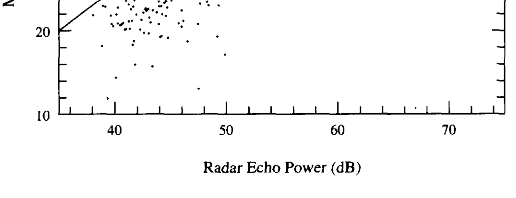

Fig. 2. Scatter plot of hourly-averaged echo power against the power predicted from the theoretical model, both expressed as dB. The model power has not been calibrated absolutely, so relative variations only are significant; this means that the mean offset between the two datasets is not significant. Also shown is a line with slope 1, which should be consistent with the data if the model were correct.

sondes travel eastward over the Welsh hills where the temperature profile will be affected by lee waves, and some of the discrepancy may be due to the different volumes sampled by the sonde and the radar, especially where peaks and troughs in power were slightly displaced in height between the two profiles. Some may be also be due to variations in the empirical parameter $F(\lambda)$. Further studies using lidar-derived temperature profiles (Vaughan et al., 1993) coincident with the radar scattering volume are planned to investigate this point.

The data were carefully examined for a scale-height dependence in the $F(\lambda)$ parameter (i.e., eq. (4)). None was found. An example of measured power plotted together with model values calculated from the best-fit straight line (i.e., slope 0.80) is shown in Fig. 1, which is typical of the data set used. Broad features in the profiles correspond very well, with some agreement also for the smaller-scale features. In this case, the two profiles agree reasonably well below 8 km and above 14 km, where the data were not used for the regression analysis. The agreement above 14 km was common to all the profiles studied, but that below 8 km is untypically good: peaks in power were generally found to coincide in altitude in this region but the absolute power level could differ by up to 10 dB. This is understandable, given the influence of water vapour on the refractive index at these heights.

3. Tropopause height

The tropopause was derived from the radiosonde profiles using the standard WMO criterion, based on the lapse rate, and compared with two parameters derived from radar data as follows: (i) the height h_{min} of the minimum echo power p_{min} measured between 5 and 11 km was noted; (ii) the height of the maximum echo power p_{max} measured above h_{min} gave the first parameter, h_{max}; (iii) the height of the first minimum in the echo power profile below h_{max} was denoted h_{min1}, provided the power $P_{max} - P_{min\,1} > 6$ dB; (iv) The height of the maximum echo power gradient between h_{max} and $h_{min\,1}$ gave the second parameter h_{grad}.

The tropopause should correspond to the maximum power gradient, as explained in Section 1. However, h_{max} is easily derived for a profile and is insensitive to small-scale structure in it. h_{grad}, despite being the most valid definition theoretically, relies on differentiating the echo power profile and is therefore more sensitive to the smaller scales, and to noise.

The results are shown in Figs. 3, 4 as scatter plots, with the sharpness of the radiosonde tropopause categorised following Price and Vaughan (1993). The data contained one anomalous profile, that for 26 March 1992. This showed a very indefinite tropopause region where the WMO definition placed the tropopause much too high, the radar tropopause was, in fact, more consistent with the humidity and ozone profiles and coin-

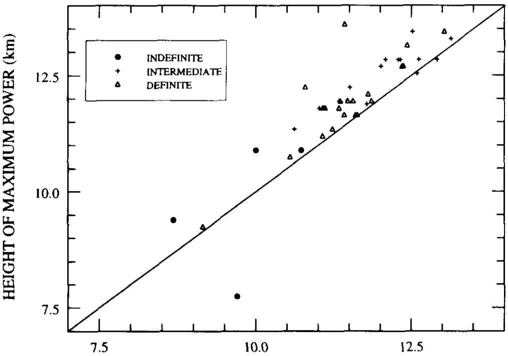

Fig. 3. Scatter plot of the height of the maximum vertical echo power, h_{max}, against radiosonde tropopause height, with the tropopause sharpness categorised according to Price and Vaughan (1993). The line shown denotes equality of the two quantities. The indefinite profile falling below the line corresponds to 26 March 1992.

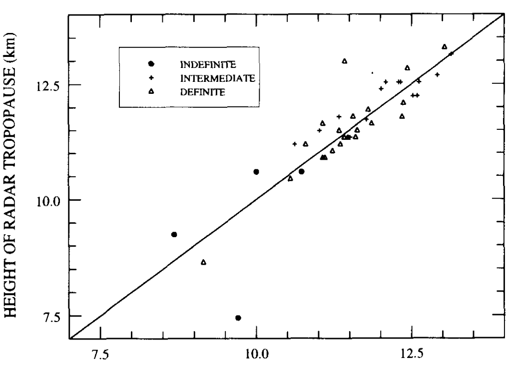

Fig. 4. As Fig. 3 except for h_{grad}.

Table 1. *Results of linear regression between the height of the tropopause determined by radar (using two different methods) and by radiosondes*

Parameter	Correlation coefficient	Mean difference	SD of diffs.
h_{max}	0.90	450 m	450 m
h_{grad}	0.91	90 m	400 m

cided with a small temperature inversion. When this point was discarded from the dataset, linear regression between the WMO tropopause and the radar parameters yielded results as shown in Table 1.

Although the mean difference between h_{grad} and the thermal tropopause is much smaller than for h_{max}, the scatter in the data and the correlation coefficients are similar in the two cases, so that h_{max}, with a suitable offset, appears to be as reliable an estimator of the tropopause position as h_{grad}. Previous studies of this kind (Gage and Green, 1979; Gage et al., 1986) using radars with lower height resolution (see above) have found similar r.m.s. variations between radar and radiosonde tropopauses, suggesting that the difference is inherent to the two techniques and not dependent on the resolution of the radar (provided this is better than 1 km). The difference probably arises from the arbitrary nature of the tropopause definition (by either method), the tropopause simply does not correspond to a physical surface. Use of ozone mixing ratio as a surrogate for potential vorticity (the favoured dynamical indicator of the tropopause, Hoskins 1991) shows even larger discrepancies with the WMO definition (Jones et al., 1994).

4. Tropopause sharpness

One feature of very indefinite tropopauses is that the static stability never approaches the dry adiabatic lapse rate in the upper troposphere. As a consequence, the echo power does not fall to a deep minimum below the tropopause as is found on a normal sounding (such as Fig. 1). This offers the possibility that the minimum echo power P_{min} may be used as an indicator of very indefinite tropopauses. There is some suggestion of this in the dataset used here (Fig. 5), where the larger

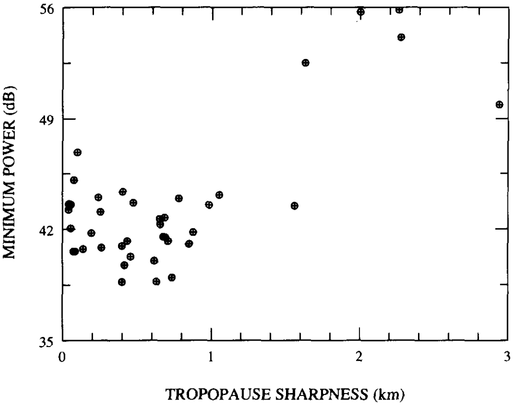

Fig. 5. Scatter plot of minimum echo power against tropopause sharpness. The latter is defined as the vertical displacement between the thermal tropopause and the highest tropospheric altitude where the lapse rate exceeds 5 K km^{-1}.

values of P_{min} clearly correspond to tropopause sharpness >2 km. Unfortunately, the number of such profiles is too few to draw a sound conclusion.

Some further evidence is provided by comparing radiosonde profiles measured from Aberporth (40 km from Aberystwyth) with data from 1–14 July 1991, when the radar ran continuously. Table 2 shows the correspondence of tropopause sharpness category and P_{min} measured at Aberystwyth during this period. The indefinite tropopauses observed 1–2 July correspond to the

Table 2. *Summary of tropopause structure and minimum radar echo power for July 1991*

Profile date/time	Minimum power	tropopause sharpness category
2 July 0600	55	indefinite
2 July 1400	53	indefinite
3 July 0600	47.5	definite
3 July 1400	45.5	definite
4 July 0600	40	definite
5 July 0600	43.5	intermediate
5 July 1400	40	definite
8 July 0600	45	intermediate
9 July 0600	47	definite
10 July 0600	43.5	definite
11 July 0600	47	intermediate
12 July 0600	43	definite

highest minimum power. On all other occasions the minimum power was below 50 dB and the tropopause definite or intermediate in character.

Taken together, Fig. 5 and Table 2 suggest that definite and intermediate tropopauses cannot be distinguished with this method, but that indefinite profiles are distinguished by $P_{min} > 50$ dB. This is an important result, since studies of the evolution of tropopause structure by Price and Vaughan (1993) show that the definite and intermediate types are interchangeable, whereas the highly indefinite profiles form a separate population which corresponds to favoured regions of stratosphere–troposphere exchange. Since P_{min} is a simple and reliable diagnostic of the radar data this demonstrates the potential for using the radar to identify such regions.

5. Conclusions

Correlation of vertical echo power profiles from the MST radar and model powers derived from radiosondes launched nearby shows agreement with the scattering model for radar echo powers <55 dB, although with considerable scatter. Above 55 dB, the measured echo power exceeds model predictions, suggesting either a non-linear dependence on $\bar{M}^2$ or the direct influence of turbulence. Radar data were used to identify the tropopause height by two different criteria: one empirical but insensitive to measurement error (maximum echo power) and one consistent with theory but more sensitive to noise (maximum power gradient, equivalent to maximum $\partial^2\theta/\partial z^2$). Although the latter was found to agree well on average with the WMO tropopause (mean offset 90 m) the scatter in the differences was the same as that found with the first method, suggesting that either method (with a suitable offset) could be used as a predictor of the tropopause height. These r.m.s. differences are similar to those found using lower resolution radars, suggesting that they arise, at least in part, from inherent differences in the two measurement techniques and the arbitrary nature with which the tropopause is defined, rather than simply through spatial variation. Lastly, indefinite tropopauses (where the thermal lapse rate does not approach the dry adiabatic in the upper troposphere) were found to coincide with minimum radar echo power >50 dB in the upper troposphere, suggesting that the radar may be used to identify and monitor such regions on a continuous basis.

6. Acknowledgements

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REFERENCES

Doviak, R. J. and Zrnic, D. S. 1984. *Doppler radar and weather observations*. Academic Press.

Gage, K. S. and Green, J. L. 1979. Tropopause detection by partial specular reflection with VHF radar. *Science* **203**, 1238–1239.

Gage, K. S. and Balsley, B. B. 1980. On the scattering and reflection mechanisms contributing to clear air radar echoes from the troposphere, stratosphere and mesosphere. *Radio Sci.* **15**, 243–257.

Gage, K. S., Balsley, B. B. and Green, J. L. 1981. Fresnel scattering model for the specular echoes observed by VHF radar. *Radio Sci.* **16**, 1447–1253.

Gage, K. S., Ecklund, W. L. and Balsley, B. B. 1985. A modified Fresnel Scattering model for the parameterisation of Fresnel returns. *Radio Sci.* **20**, 1393–1501.

Gage, K. S., Ecklund, W. L., Riddle, A. and Balsley, B. B. 1986. Objective tropopause height determination using low resolution VHF radar observations. *J. Atmos. Ocean. Tech.* **3**, 248–254.

Hoskins, B. J. 1991. Towards a PV-theta view of the general circulation. *Tellus* **A43**, 27–35.

Hocking, W. K. and Rottger, J. 1983. Pulse-length dependence of radar signal strengths for Fresnel backscatter. *Radio Sci.* **18**, 1312–1324.

Jones, S. B., Vaughan, G., Reid, S. J. and Smit, H. G. J. 1994. *The nature of the tropopause as revealed by ozonesondes. The life cycles of extratropical cyclones*, vol. 2, ed. S. Gronas and M. A. Shapiro. Geophysical Institute, University of Bergen, pp. 444–448.

Price, J. D. and Vaughan, G. 1993. The potential for stratosphere–troposphere exchange in cut-off low systems. *Q. J. Roy. Met. Soc.* **119**, 343–65.

Salathe, E. P. and Smith, R. B. 1992. In situ observations of temperature microstructure above and below the tropopause. *J. Atmos. Sci.* **49**, 2032–6.

Slater, K., Stevens, A. D., Pearmain, S. A. M., Eccles, D., Hall, A. J., Bennett, R. G. T., France, L., Roberts, G., Olewicz, Z. K. and Thomas, L. 1992. Overview of the MST radar system at Aberystwyth. *Proceedings of the 5th Workshop on Technical aspects of MST radar*, ed. B. Edwards, pp. 479–82. Available from the SCOSTEP secretariat, University of Illinois, 1406 W. Green St., Urbana, Illinois 61801, USA.

Vaughan, G., Wareing, D. P., Pepler, S. J., Thomas, L. and Mitev, V. 1993. Atmospheric temperature measurements by rotational Raman scattering. *Applied Optics* **32**, 2758–64.