The unexpected dominance of stratiform precipitation return power over clear-air return power for observations made at lower-VHF

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1. Introduction

Lower-VHF radars are designed to detect clear air returns, i.e. those backscattered from refractive index irregularities. Nevertheless, they are also able to detect the Rayleigh backscatter from hydrometeors if the precipitation rate is sufficiently high; the hydrometeor reflectivity increases with increasing precipitation rate. Following the theoretical considerations of *Ralph* (1995), a rain rate of between 3.2 and 56.1 mm hour^{-1} is required in order for the radar return power from precipitation to exceed that from the clear air at 46.5 MHz. The lower limit corresponds to typical peak clear air reflectivity and the upper limit to unusually strong clear air reflectivity. Consequently it is expected that hydrometeor returns should dominate under convective precipitation conditions, for which rain rates of the order of 10 mm hour^{-1} are typical and instantaneous rain rates of the order of 100 mm hour^{-1} are possible. However, clear air returns are expected to dominate under stratiform precipitation conditions, for which rain rates of the order of 1 mm hour^{-1} are typical.

This abstract relates to observations made by the 46.5 MHz MST radar at Aberystwyth. A tipping bucket rain gauge at the radar site provides 10-minute mean rain rates. Most of the precipitation falling over the UK is of the stratiform type. It tends to be related to synoptic scale features and so persists, over a given location, for several hours at a time. Although deep convection conditions are comparatively rare, they are occasionally observed by the Aberystwyth radar (e.g. *Hooper et al.*, 2005). They have typical transit times over the radar site of the order of a few tens of minutes.

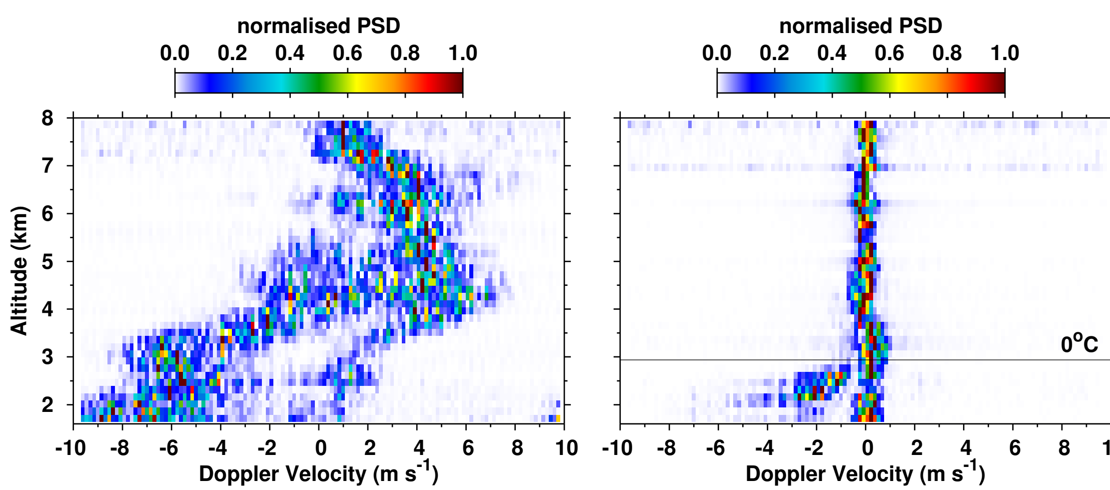

Figure 1: Radar return spectra for vertical beam observations made (left panel) at 13:23 UT on 1st March 2003, under conditions of convective precipitation, and (right panel) at 08:00 UT on 28th May 2004, under conditions of stratiform precipitation.

2. Direct observations of hydrometeor returns

The spectra shown in the left panel of Figure 1 relate to a convective event, during which the 10-minute mean rain rate was 19.2 mm hour^{-1}. Two sources of radar returns are apparent at altitudes below 5 km. The clear air returns are those with the more positive Doppler velocities, where a positive value implies upward motion. Updrafts of up to 10 m s^{-1} are a typical feature of convective events observed by the Aberystwyth radar. The hydrometeor returns are those with the more negative Doppler velocities. Terminal fall speeds of up to 10 m s^{-1} are also typical. As might be expected, the peak power spectral densities (PSDs) of the hydrometeor returns exceed those of the clear air returns over the entire observed altitude range.

The right panel of Figure 1 relates to 08:00 UT on 28th May 2004, during a stratiform precipitation event which lasted from 06:10 until 11:30 UT; see the bottom panel of Figure 2. The tipping bucket rain gauge records a tip for each 0.2 mm accumulation of rain. Consequently 10-minute mean rates are integer multiples of 1.2 mm hour^{-1} and little significance should be attached to the fluctuations seen in the bottom panel of Figure 2. These are simply indications of an actual rain rate of close to minimum detectable 1.2 mm hour^{-1}. The appearance of hydrometeor echoes with increasingly negative Doppler velocities with decreasing altitude beneath the 0^{o}C level is to be expected (the temperature is taken from a Camborne radiosonde, which is described in more detail below). This is caused by the melting of snow flakes and the subsequent acceleration of the rain drops; rain drops have a higher reflectivity than snow flakes. What is unexpected is that the peak PSDs of the hydrometeor returns should exceed those of clear air returns for such a low rain rate. Nevertheless this is clearly the case between altitudes of 2 and 3 km and it is commonly observed during stratiform precipitation events.

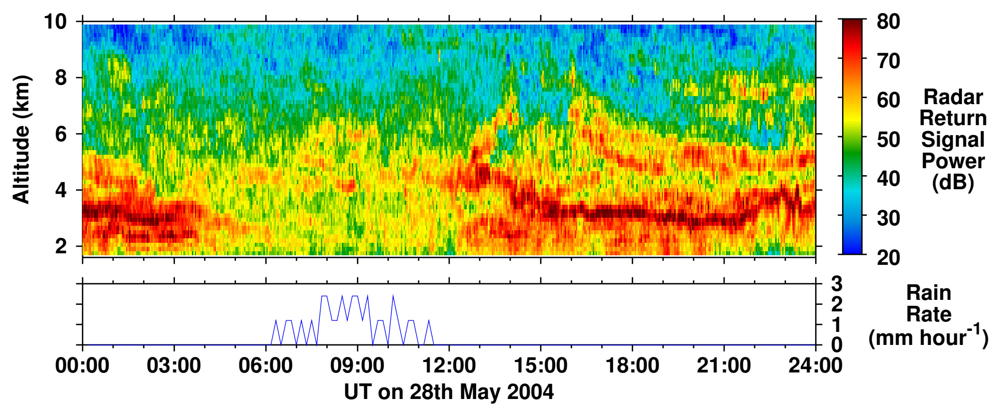

Figure 2: (top panel) MST radar return signal power and (bottom panel) 10-minute mean rain rate observed on 28th May 2004.

3. A quantitative analysis of the clear air return power

The primary reason for this unexpected result is immediately apparent from the top panel of Figure 2. It is well-known that precipitation can lead to a suppression of the clear air radar return power (e.g. *McDonald et al.*, 2006). In this case the suppression, at an altitude of 3 km, has a maximum value of 20 dB. Moreover the hydrometeor reflectivity peaks in a bright band at the melting layer, as seen at higher radar frequencies (e.g. *Ralph*, 1995). Under such conditions the radar return power from hydrometeors can dominate at a lower rain rate than suggested by the theoretical considerations of *Ralph* (1995). A rain rate of the order of 1 mm hour^{-1} is clearly sufficient under these circumstances.

For the majority of conditions, the clear air radar return power, for a vertically directed beam, is found to be proportional to M^2/z^2, where M is the mean vertical gradient of generalised potential refractive index and z (m) is the altitude above the radar (e.g. *Hooper et al.*, 2005). The dry component of M is given by M_D:

$$M_D = -77.6 \times 10^{-6} \frac{p}{T} \frac{\partial \ln \theta}{\partial z} \quad (1)$$

where p (hPa) is pressure, T (K) is absolute temperature and θ (K) is potential temperature. M_D is closely related to the square of the Brunt-Väisälä frequency, [$\omega_B^2 = g(\partial \ln \theta)/\partial z)$ rad^2 s^{-2}]. The full term for M additionally depends on the both the specific humidity, q (kg kg^{-1}) and its vertical gradient:

$$M = M_D \left[1 + \frac{15500q}{T} - \frac{\frac{7800}{T}\frac{\partial q}{\partial z}}{\frac{\partial \ln \theta}{\partial z}} \right] \quad (2)$$

Temperature, pressure and relative humidity data are taken from the radiosonde launched at 11:16 UT on 28th May 2004 from Camborne, which is located approximately 270 km to the south of Aberystwyth. Profiles of the square of the Brunt-Väisälä frequency and of the relative humidity are shown in the left and middle panels, respectively, of Figure 3. The dry and full predictions of clear air radar return power based on these data are shown by the thin and thick yellow lines, respectively, in the right panel of Figure 3. Attention is drawn to the fact that the humidity terms are responsible for an enhancement of up to 20 dB over the dry predictions at altitudes below 4 km. There is a good general agreement between the median observed radar return power profile for the period 00 - 03 UT (i.e. before the onset of the precipitation), as indicated by the thick blue line, and the full prediction. Both the radar and radiosonde sites are located within the same air mass at the time of this sounding. Consequently, despite the large separation between them, the radiosonde data are clearly broadly representative of the atmosphere above Aberystwyth. The lack of agreement between altitudes of 6 and 9 km suggests that the humidity feature encountered by the radiosonde is not a synoptic-scale feature.

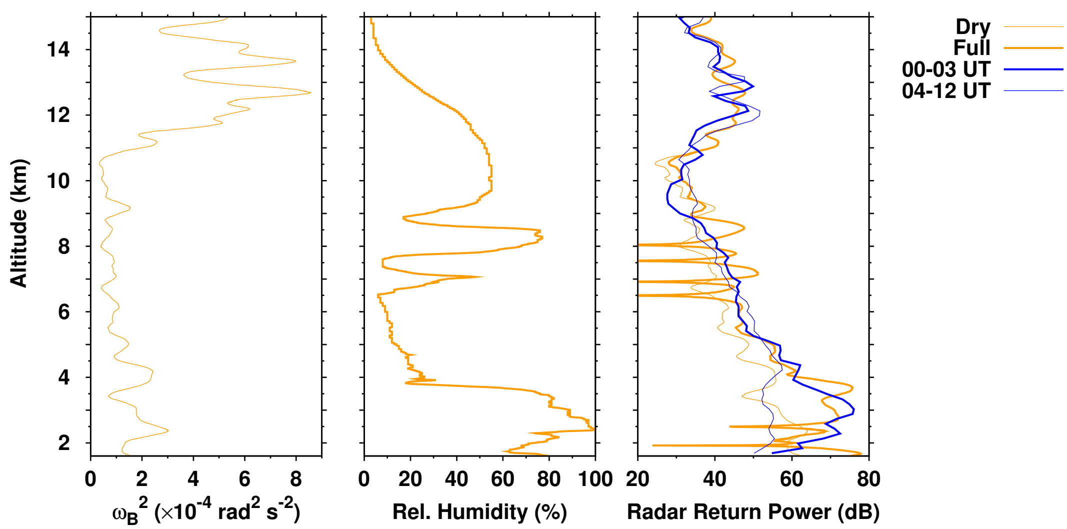

Figure 3: (left panel) Square of the Brunt-Väisälä frequency and (middle panel) relative humidity for a radiosonde launched from Camborne at 11:16 UT on 28th May 2004. (right panel) The clear air radar signal power predicted using the dry and full terms are shown together with the median observed powers for the periods 00 - 03 UT and 04 - 12 UT.

Although rain is only seen at the ground between 06:10 and 11:30 UT, the spectral data (not shown) indicate that precipitation is occurring aloft outside of these times. Clearly the rain is evaporating before it reaches the ground (the peak PSDs of the hydrometeor returns do not always exceed the peak PSDs of the clear air returns at times when rain does not reach the ground). The median power profile representing the precipitation period is consequently taken over the interval 04 - 12 UT, indicated by the thin blue line. It is apparent from Figure 3 that the above model of clear air radar return signal power is not appropriate under precipitation conditions. It is noted, however, that the dry prediction is in reasonable agreement with the observed values at the altitudes where the clear air return power is suppressed by the precipitation, i.e. below 4 km.

On the basis of these results, changes have been made to the MST radar signal processing scheme for the Aberystwyth radar. In the cases of both convective and stratiform precipitation, the clear air and hydrometeor returns tend to partially overlap. The PSDs in the white regions between the signal components shown in Figure 1 are significantly above the noise PSD. An earlier method of identifying signal limits followed a smoothed PSD envelope down from either side of the peak (smoothed) PSD until the noise PSD was reached, or until the smoothed PSD dropped below a hundredth of the peak value. The latter condition was found to be effective at preventing weak PSDs at the extremities of the signal limits from having a disproportionate influence on the spectral width. However, these conditions lead to both signal components being identified as part of a single, broad signal component. The resulting signal parameters are consequently representative of neither the clear air returns nor the hydrometeor returns. Although this is less of a problem for convective events, since they persist for such short time scales, stratiform precipitation tends to persist for several hours at a time. This leads to contamination of the wind-profile data.

An improved scheme additionally imposes a signal limit where a local minimum of the smoothed PSD envelope exists at less than a tenth of the peak smoothed PSD value. This simple measure has been found to be highly effective. It is recognised that more sophisticated methods exist for resolving partially overlapping signal components (e.g. *Boyer et al.*, 2004). There are no plans to introduce such techniques into the operational signal processing scheme for the Aberystwyth MST radar in the immediate future.

References

Boyer, E., M. Petitdidier, and P. Larzabal, Stochastic Maximum Likelihood (SML) parametric estimation of overlapped doppler echoes, *Ann. Geophys.*, *22*, 3983–3993, 2004.

Hooper, D. A., A. J. McDonald, E. Pavelin, T. K. Carey-Smith, and C. L. Pascoe, The signature of mid-latitude convection observed by VHF wind-profiling radar, *Geophys. Res. Lett.*, *32*, L04808, doi:10.1029/2004GL020,401, 2005.

McDonald, A. J., K. P. Monahan, D. A. Hooper, and C. T. Gaffard, VHF signal power suppression in stratiform and convective precipitation, *Ann. Geophys.*, *24*, 23–35, 2006.

Ralph, F. M., Using radar-measured radial vertical velocities to distinguish precipitation scattering from clear-air scattering, *J. Atmos. Ocean. Tech.*, *12*, 257–267, 1995.