4A.5 DERIVING TURBULENT KINETIC ENERGY DISSIPATION RATE WITHIN CLOUDS USING GROUND BASED 94 GHZ RADAR

Dominique Bouniol*, Anthony J. Illingworth, Robin J. Hogan

Centre d'étude des Environnemant Terrestre et Planétaire, Vélizy, France

University of Reading, Reading, United Kingdom

1. INTRODUCTION

Small-scale turbulence has an important influence on the cloud structure and therefore on the spatial distribution of the optical and microphysics properties. Turbulence is also directly linked to the life cycle of the clouds through internal mixing and entrainment processes. A good way to estimate the turbulence activity is to measure the turbulent kinetic energy (TKE) dissipation rate (ε) which represents the rate of conversion of TKE into heat or in other words the rate at which the TKE is dissipated by viscosity.

Several methods have been proposed in order to derive this parameter from observations. The most common one is to perform spectral analysis on aircraft data (for instance Gultepe and Starr (1995)) or from ground based radar data (Brewster and Zrnić 1986). Another technique is to derive ε from ground based spectral width measurements such as Chapman and Browning (2001) or Kollias *et al.* (2001). Turbulence inferences made at vertical incidence such as those by Kollias *et al.* (2001) may be biased, especially at low turbulence levels, because of the contribution from the spread of terminal velocities in the particle size distribution.

In this paper a new method, inspired by the early work of Rogers and Tripp (1964), is presented. It consists of exploiting the very high temporal resolution of Doppler measurements performed by ground based 94 GHz radar in order to derive ε within non-precipitating clouds which can be composed of water droplets such as stratocumulus or ice crystals such as cirrus.

As a first step the principle of the method is explained, then the underlying hypotheses are tested and finally the method is applied to data collected by the 94 GHz radar of Chilbolton (UK) operated in the framework of European Project CloudNet.

2. PRINCIPLE

The first hypothesis which must be justified is that the radar targets (*e.g.* water droplets or ice crystals) are good tracers of the turbulent motions; this is why this method can only be applied to non-precipitating ice clouds or to particular water clouds such as stratocumulus which contain only small drizzle droplets. As a second step the TKE spectrum is assumed to be Kolmogorov in form which means that turbulence is isotropic and homogeneous at the scale sampled by the radar. Figure 1 displays the shape of the TKE spectrum and shows that within the inertial subrange a relationship should exist between the large-scale velocity fluctuations and the TKE dissipation which occurs at wavelengths of a centimeter or less.

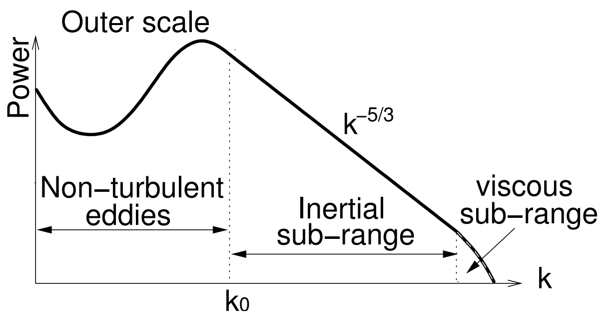

FIGURE 1: Schematic plot of the TKE spectrum. The dotted line indicates the different regions of the TKE spectrum which are labelled.

Kolmogorov's theory states that, within the inertial sub-range, the spectral density can be expressed as:

$$S(k) = a\varepsilon^{2/3}k^{-5/3}$$

where a is the universal Kolmogorov constant with a value of 1.62 and k is the wave number.

Rogers and Tripp (1964) show that the average TKE per unit mass of air (E) can be written as:

$$\sigma_{\overline{V}}^2 + \overline{\sigma_t^2} = 2E$$

where $\sigma_{\overline{V}}^2$ is the variance of the mean Doppler velocity and σ_t^2 is the variance due to turbulence within the pulse volume. In addition to having the property that their sum is proportional to the TKE the two variances are such that their relative magnitudes indicate the manner in which the TKE is partitioned between different scales. The variance

*Corresponding author address: Dominique Bouniol, CETP/CNRS, 10/12 avenue de l'Europe, 78140 Vélizy, France; e-mail: dominique.bouniol@cetp.ipsl.fr

$\sigma_{\overline{v}}^2$ of the mean wind is an indicator of the kinetic energy in turbulent scales that are generally larger than the dimensions of the sampled volume. Conversely the average variance $\overline{\sigma_t^2}$ is an indicator of the kinetic energy energy in the small scales of turbulence (Rogers and Tripp (1964)).

These variances can be thought of as part of the TKE spectrum. The variance due to turbulence within the pulse volume may be rewritten as follows:

$$\sigma_t^2 = \int_{k_1}^{k_2} a\varepsilon^{2/3}k^{-5/3}dk$$

where $k_2 = 2\pi/L_2 = \lambda/2$ is the smallest scale that can be probed by the Doppler radar, ultimately $\lambda/2$ (where λ is the radar wavelength), $k_1 = 2\pi/L_1$ is related to the scattering volume dimension and so includes large eddies traveling through the sampling volume during the dwell time (Kollias *et al.* 2001). This variance contributes directly to the spectral width and must be extracted from the other contributions to spectral width, such as wind shear and particle terminal fall speed, if we want to estimate ε from this measurement (see for instance Chapman and Browning (2001) or Kollias *et al.* (2001)).

In the same way the variance of the mean Doppler from successive spectra can also be expressed:

$$\sigma_{\overline{v}}^2 = \int_{k}^{k_1} a\varepsilon^{2/3}k^{-5/3}dk$$

where $k = 2\pi/L$ is related to the large eddies travelling through the sampling volume during the sampling time (typically about 30 s). The different L parameters represent the length scales related to each wave number.

By integrating this last expression and assuming that $k_1^{-1} \ll k^{-1}$ ε can be expressed as:

$$\varepsilon = \left(\frac{2}{3a}\right)^{3/2} \sigma_{\overline{v}}^3 k$$

The main contribution to k is due to the advection of clouds by the wind through the beam, a distance which is generally much larger than the width of the radar beam at a given altitude. In this paper the horizontal wind is derived from the ECMWF analysis (stored every hour at the same geographical location in the framework of the CloudNet project). The wave number can then be written as:

$$k = \frac{2\pi}{L} = \frac{2\pi}{\text{beamwidth} + T_s\|\overrightarrow{V_h}\|}$$

where T_s is the sampling time, $\|\overrightarrow{V_h}\|$ the modulus of the horizontal wind interpolated from the hourly analysis and the beamwidth can be computed at a given altitude by using beamwidth = $2z\sin(\theta/2)$ with $\theta=0.5^\circ$ in our case.

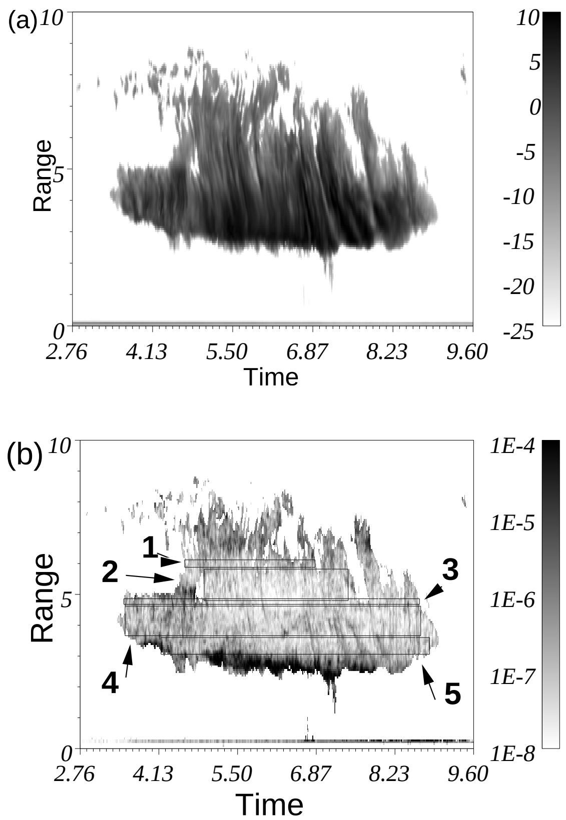

FIGURE 2: (a) Reflectivity in [dBZ] observed by the vertically pointing 94 GHz Doppler radar at Chilbolton (UK) on the 17th July 2001. (b) Estimated ε in $[m^2.s^{-3}]$ by using the previous explained method. The boxes displayed within the cloud and labelled from 1 to 5 are used for validation in the next section.

As an illustration the method is applied to a deep ice cloud observed on the 17th July 2001. Figure 2(a) shows the observed reflectivity which is in the range -25 dBZ to 10 dBZ. Lower values than -25 dBZ are not detected at this time due to a loss of sensitivity of the radar at this time (see Hogan *et al.* 2003). Figure 2(b) shows the values of ε which range from 10^{-4} $m^2.s^{-3}$ at the base of the cloud to 10^{-8} $m^2.s^{-3}$ within the interior of the cloud. These values are in the same range as those obtained in previous studies (see for instance the summary table of Gultepe and Starr (1995)).

These initial values seem consistent with those reported in the literature. We now present a detailed analysis to justify the assumptions made in this radar technique.

3. VALIDATION

In this section we evaluate the two hypotheses on which the radar method is based. First we have to be sure that the TKE spectrum is of the Kolmogorov form. To test the first hypothesis we show that the value of ε is independent of the number of one second mean Doppler velocities used to compute $\sigma_{\overline{V}}^2$ thus demonstrating that we are staying within the inertial subrange. The mean value of ε within each of the areas labelled in Fig. 2(b) is computed for different numbers of mean Doppler estimates and the results are displayed in Fig. 3.

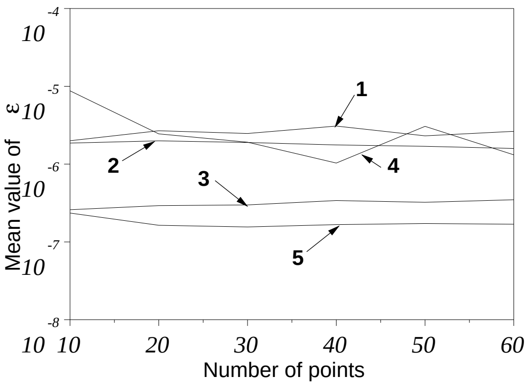

FIGURE 3: Mean value of ε in [$m^2.s^{-3}$] for the areas shown on Fig. 2(b) computed for different number of points.

Figure 3 shows that the mean values of ε are not sensitive to the number of points used to compute the $\sigma_{\overline{V}}^2$. A slight variation in the results can be observed for the 10 and 20 second dwells, however for the longer dwells the mean value of ε is nearly constant. This is mainly due to the fact that for computing ε, $\sigma_{\overline{V}}^2$ is multiplied by k in which appears the sampling time which is itself related to the number of points. What can also be observed from Fig. 3 is that even though the mean is computed over 60 s it is still constant, and that on this scale there is no contribution from any variations in the terminal velocity which could result if patches of precipitation were drifting past.

As a second step to justify this method based of computating $\sigma_{\overline{V}}^2$ in order to retrieve ε, we now compare it with the classical way of estimating ε by using the spectral analysis of the long time series. In this case, ε values are computed using 30 points and then a FFT is applied to each 30 point sample. Then the spectrum is fitted (only for points above the noise level) with a power law of the type $\alpha k^{-5/3}$. The value obtained for α corresponds to $a\varepsilon^{2/3}$.

The comparison of individual values of ε by the two methods is difficult since the FFTs are computed using a small number of points so a statistical comparison is more representative. Figure 4 shows a scatter plot of the ε values obtained from the FFT against the computation of $\sigma_{\overline{V}}^2$. The two estimators are in very good agreement, the data are spread closely around the 1/1 line (the solid black line in Fig. 4). However it is observed that the spread does increase for the smallest values of ε especially for ε lower than 10^{-7} $m^2.s^{-3}$. Testing this method of estimating ε against a spectral analysis is in fact equivalent to checking the second hypothesis performed in the previous section: ($k_1^{-1} \ll k^{-1}$).

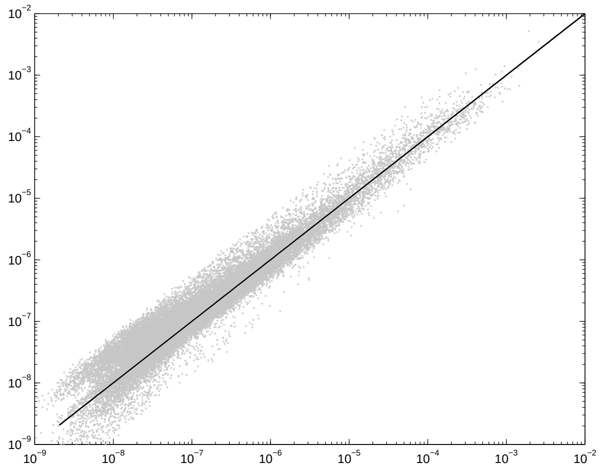

FIGURE 4: Scatter plot of value of ε in [$m^2.s^{-3}$] with ε values deduced from the present analysis on the x-axis and ε deduced from spectral analysis on the y-axis.

In this section, the main hypothesis (*i.e.* that the TKE spectrum is Kolmogorov in form) has been evaluated by using different times to compute $\sigma_{\overline{V}}^2$. The deduced values of ε are not sensitive to the sampling time, showing that the Kolmogorov theory is a good approximation for TKE spectrum for our purpose and that the time interval typically used (about 30 s) stays within the inertial subrange. A point to point comparison between this analysis and a spectral analysis has been carried out and also shows a very good agreement between the two ways of estimating ε. We conclude that the use of the variance of the mean Doppler provides a reliable estimate of ε.

4. RESULTS

One of the main strengths of this approach is that ε can be evaluated for a large data set of radar measurements, provided that the time resolution is such that the data are within the inertial sub-range and that an estimate of the horizontal wind is available. It is then possible to investigate if some common characteristics exist for the dissipation of TKE within different kind of clouds and to see if these characteristics are related to the microphysical, radiative and dynamical processes in the clouds. This investigation is possible since a large data set is available from the 1 year of data obtained in the framework of CloudNet.

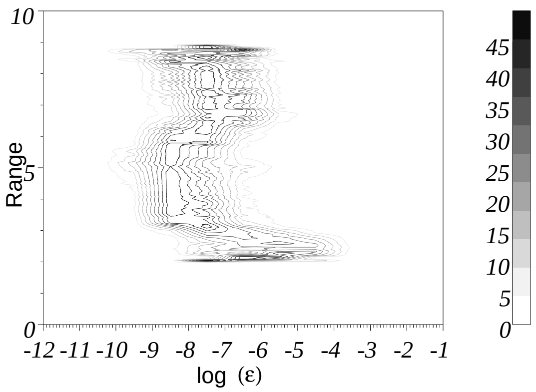

FIGURE 5: Histogram of log(ε) in [$m^2.s^{-3}$] by altitude in [km] for the 17th July 2001. Contour levels correspond to percentage of occurrence of a given value of log ε at each altitude.

This technique is being applied to the whole data set and several characteristics can be identified. For instance, for the case of the 17th July 2001, the histogram of the frequency of the values of ε as a function of altitude displayed in Fig. 5 and shows the typical features often encountered within ice cloud: high values at the top and even larger values at the bottom but much smaller values in the interior. Similar behaviour was found by Hogan and Illingworth (2003) in their study of the horizontal variability of ice water content as a function of the distance from cloud top and base. The increase at cloud base is probably due to turbulence resulting from the evaporative cooling where the ice particles fall in an unsaturated environment, The generality of this kind of behaviour is under investigation and more detailed results will be give during the presentation.

Another characteristic often encountered within pre-frontal clouds are thin layers having a higher level of turbulence than their surroundings. For instance the typical values of ε observed within the cloud core are like shown in Fig. 5 about 10^{-8}–10^{-7} and these thin layers present values about 10^{-5} *i.e.* two or three order of magnitude larger. These regions appear to coincide with localised intense vertical wind shear.

Analysis of stratocumulus clouds reveals in-cloud values within the range 10^{-4}–10^{-2} in agreement with previous studies. No particular correlations with the stratocumulus dynamics or microphysics is obvious. However a common characteristic is that within drizzle, identified by a radar cloud base extending below the lidar cloud base, the turbulence activity is one order of magnitude smaller (about 10^{-5}).

Further statistical analysis of the whole data set will be presented at the conference.

ACKNOWLEDGEMENTS

This work was carried out with the support of EU CloudNet contract EVK2-CT-2000-00065 and NERC grant NER/T/S/1999/00105.

5. REFERENCES

Brewster, K. A. and D. S. Zrnić, 1986: Comparison of Eddy Dissipation Rates from Spatial Spectra of Doppler Velocities and Doppler Spectrum Widths. *J. Atmos. Oceanic Technol*, **3**, 440–452.

Chapman, D. and K. A. Browning, 2001: Measurements of dissipation rate in frontal zones. *Quart. J. Roy. Meteor. Soc.*, **122**, 1939–1960.

Kollias, P., Albrecht, B. A., Lhermitte, R. and A. Savtchenko, 2001:Radar Observations of Updrafts, Downdrafts, and Turbulence in Fair-Weather Cumuli. *J. Atmos. Sci.*, **58**, 1750–1766.

Gultepe, I. and D. O'C. Starr, 1995: Dynamical Structure and Turbulence in Cirrus Clouds: Aircraft Observations during FIRE. *J. Atmos. Sci*, **52**, 4159–4182.

Hogan, R. J., D. Bouniol, D. N. Ladd, E. J. O'Connor and A. J. Illingworth, 2003 : Absolute calibration of 94/95 GHz radars using rain. *J. Atmos. Oceanic Technol.*, **20**, 572–580.

Hogan, R. J. and A. J.Illingworth, 2003: Parameterizing ice cloud inhomogeneity and the overlap of inhomogeneities using cloud radar data. *J. Atmos. Sci.,* **60**, 756–767.

Rogers, R. R. and B. R. Tripp, 1964: Some Radar Measurements of Turbulence in Snow. *J. Appl. Meteor.*, **3**, 603–610.