Fine Altitude Resolution Observations of Stratospheric Turbulent Layers by the Arecibo 430 MHz Radar

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(Manuscript received 28 July 1981, in final form 26 July 1982)

ABSTRACT

Stratospheric turbulence is now observed by the Arecibo 430 MHz radar with an improved altitude resolution of 150 m. Turbulence is observed to occur in thin layers with a thickness usually less than the altitude resolution, and estimated to be about 50 m. A clear correlation is found between the power, wind shear and spectral width of the echoes. A simple method of estimating the energy dissipation rate and the eddy diffusivity is examined. Values of the order of 0.2 $m^2 s^{-1}$ are found for the eddy diffusivity coefficient in the lower stratosphere, showing the importance of turbulence on vertical transport.

1. Introduction

At the time of the first stratospheric experiment at Jicamarca, Peru (Woodman and Guillen, 1974), it had already been pointed out that the stratospheric turbulent layers must be a few tens of meters thick. This has been proven by recent observations with higher altitude resolution (Rottger and Schmidt, 1979; Woodman *et al.*, 1980; Woodman 1980a,b).

This paper briefly presents the basic characteristics of the stratospheric turbulent layers found by the Arecibo 430 MHz radar through a series of high resolution observations carried out since 1978. It will be shown that the present altitude resolution of 150 m is still not sufficient to resolve the microstructure of the turbulent layers, but seems to be enough to separate individual scattering layers, avoiding the problem of multiple scattering layers in one height sample.

One of the most important objectives in the radar study of the stratosphere is to understand the role of stratospheric turbulence in vertical transport phenomena. Various techniques have been applied to measure the energy dissipation rate and the eddy diffusivity, and have resulted in a rather different conclusion on the importance of stratospheric turbulence (Woodman, 1981). In this paper, an attempt is made using a simple method to compute these turbulence parameters from the observed width of the echo power spectra.

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0022-4928/82/112546-07$05.75

2. Experiment

The details of the experimental technique and the analysis procedure for determining the basic physical parameters are reported elsewhere (Woodman, 1980a; Sato and Woodman, 1982b). The antenna beam was switched every one hour or so, between the eastward and northward directions at 7.5–15° zenith angle. A 32-bit complementary coded pulse of 1 μs basic pulse length was transmitted every 730 μs at 2 MW peak power. The received echo was sampled at 256 heights at 150 m spacing, beginning from the ground level. Meaningful data was obtained from the 5–30 km height range only. The lower level was determined by receiver cutoff and the upper by system sensitivity. The radar returns were processed on real time in terms of power frequency spectra using an array processor, integrated one or two minutes and then stored on magnetic tapes for further off-line processing.

A nonlinear parameter estimation procedure was applied to this data in order to remove the effect of a very strong fading ground clutter echo, and to determine the first three spectral moments of the signal at the same time (Sato and Woodman, 1982b).

3. Results

Fig. 1 is an example of an altitude profile of the scattered echo power. The large fluctuation of the power with height indicates that the scatterers are not distributed uniformly, but rather concentrated in discrete layers. The power decreases a few to 10 dB at adjacent heights around most maxima. Since the range gates are spaced at the nominal altitude resolution of 150 m, this sharp decrease in power around

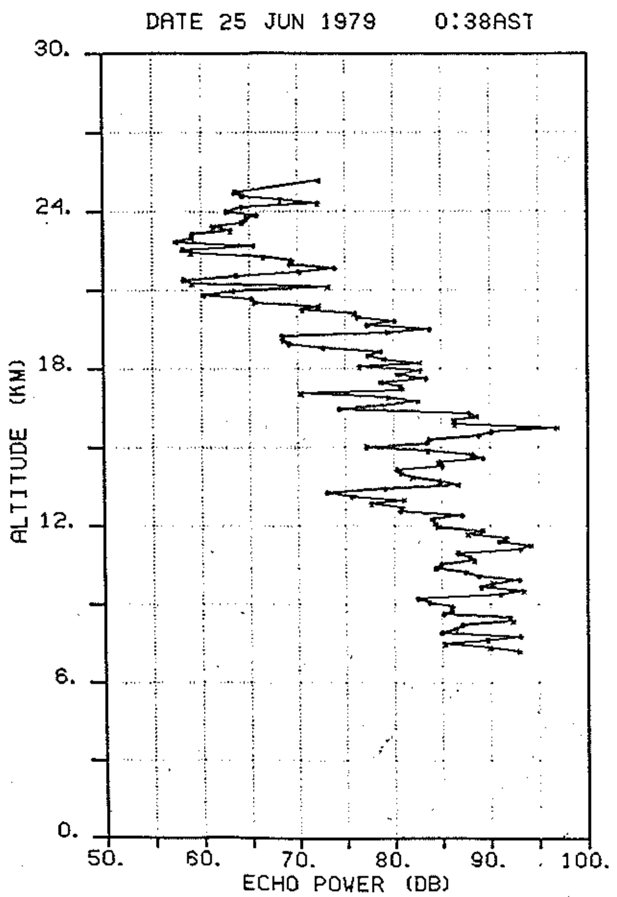

FIG. 1. The received echo power versus height (after Sato and Woodman, 1980). The noise level is about 59 dB.

the peaks means that most layers are thinner than 150 m.

Much better altitude resolution is required to observe microstructures of the layer directly. However, the sharp minima in power between strong layers indicates that this altitude resolution is sufficient to separate individual turbulent layers. Vertical separation between layers ranges from a few hundred meters to one–two kilometers.

Fig. 2 shows a height–time variation of the echo power in an 8-level shade plot. The dynamic range from white to black corresponds to 32 dB. Because of this, unfortunately, the contrast is low and the layered structure does not show as dramatically as in Fig. 1. The antenna beam was pointed east and north at 15° zenith angle alternately during every one hour interval. A fairly good continuity between the power in these two directions indicates the uniformity of the turbulent layers over 3–10 km in the horizontal direction.

A descending layer which starts from 13 km at the beginning of Fig. 2 is a typical example of turbulent layers in the troposphere. If we associate this layer with a point of maximum wind shear, this movement would seem to indicate the passage of a solitary gravity wave propagating its energy upward. It is important that these thin layers show vertical motion because the slowly sweeping motion of a turbulent layer increases its effective diffusivity several times more than a layer with the same thickness (Woodman *et al.*, 1981) which stays at the same altitude. The scattering layers in the stratosphere (above 16 km) are more stable than those in the troposphere; close inspection, nevertheless, show the same sweeping motion, but through a smaller altitude range. The layers preserve their own identity sometimes for several hours. Since the layers should be moving at the speed of the mean flow, this continuity in time means the horizontal extent of these layers is on the order of 100 km. It is surprising that such extensive layers have a thickness of 150 m or less. This stationarity is, however, consistent with the observed stability of the vertical shear structure of the horizontal velocity (Sato and Woodman, 1982a).

The Richardson number calculated at the two strongest peaks at 15.5 and 17 km using balloon temperature data is slightly lower than 1.0, and would possibly have been even lower, had it been evaluated at altitude scales comparable to the thickness of the layer. In fact, it should be smaller, for turbulence to be energetically possible. This low Richardson number is consistent with the persistence of the strong echo at these heights throughout the experiment.

There is a good correlation between the echo power, the vertical shear of horizontal velocity, and the spectral width of the echoes. Fig. 3 shows ten-hour-average altitude profiles of these three parameters. The shear is determined by a central finite difference over 300 m. The two thin lines around the thick line indicate the standard deviation from the mean. Although the total vector should be used for this comparison, only the east–west component is shown, since only one component can be measured at a time, and the east–west component is dominant over the north–south component in this case. The three large peaks at 15–17 km in the power profile clearly have corresponding peaks in the shear and width profiles (second moment normalized with respect to power). The correspondence between shear and the other two profiles becomes worse above 18 km, probably due to the exclusion of the north–south component. Resemblance between the power and the width profiles is very good throughout the observed height range.

This positive correlation between these three parameters can be qualitatively interpreted as follows. The larger the shear, the larger the amount of kinetic energy going into the potential energy gain and random kinetic energy required by turbulence. A fluctuating temperature profile has a higher potential energy level than the stable initial stratified state. Therefore, the higher the available potential energy, the

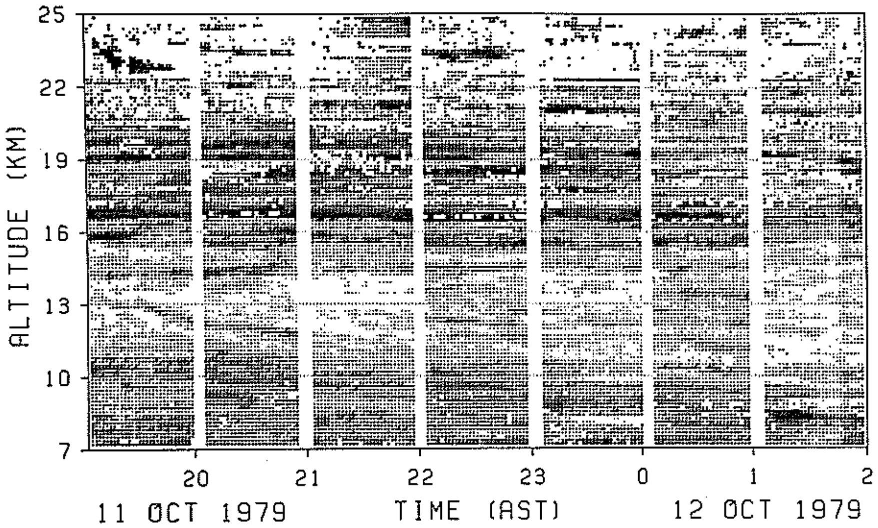

FIG. 2. 8-level height–time shade plot of the echo power. The dynamic range is 32 dB. Missing points due to poor quality of data left blank.

larger the possible temperature fluctuations and hence the larger the backscatter returns. Concurrently, the higher random turbulent kinetic energy increases the Doppler frequency spread, hence the spectral width. Shear broadening could also contribute directly to this correlation, as will be discussed later.

Since we have and will be interpreting the spectral width of the radar returns in terms of turbulent Doppler broadening, it is important to discuss one other possible source of spectral broadening, that due to finite antenna beam width. The Arecibo antenna has a diameter of 300 m with a half-beam at 430 MHz of the order of 0.1° (0.00175 radians) in the far field. In the near field, the beam is almost cylindrical with a diameter equal to that of the antenna. However, the effective beam angle, as far as finite beam broadening is concerned, is given by the angle which subtends a Fresnel zone at a given range. This can

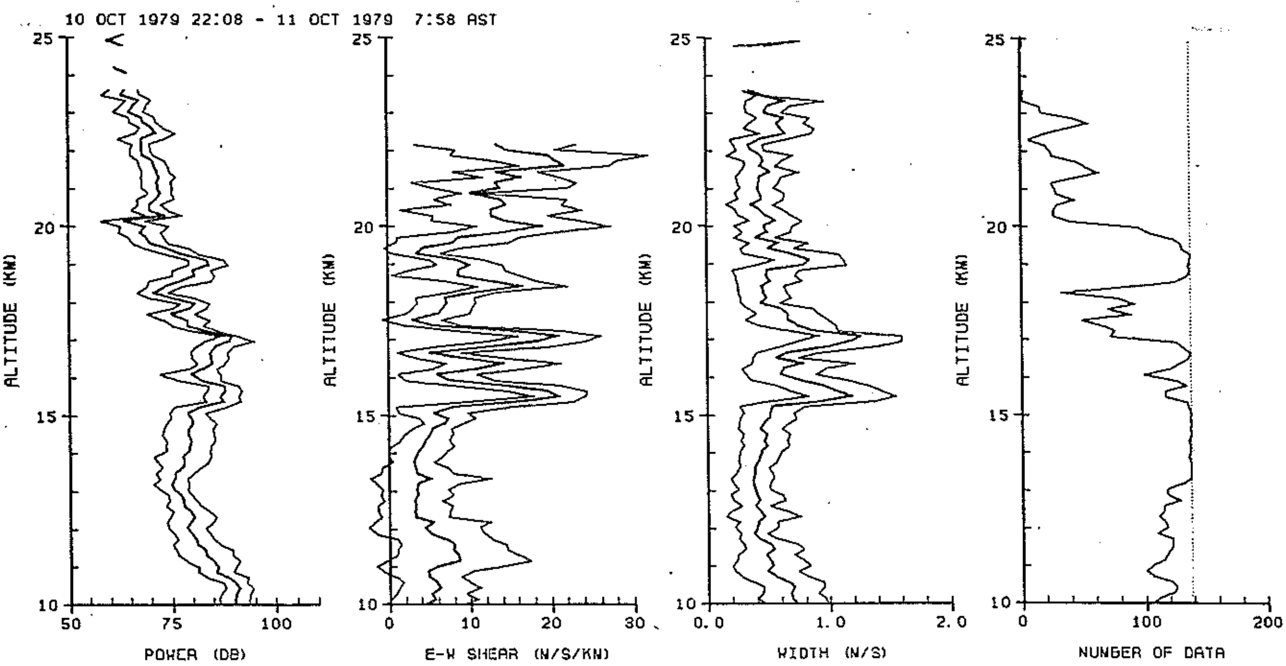

FIG. 3. Ten-hour-mean echo power, E–W component of the wind shear, and the spectral width versus height. The number of good data points used in the average are plotted on the right. The dashed line in the figure shows the total number of data points. Two thin lines around the thick line in each profile indicate the standard deviation from the mean.

be clearly seen when one considers that, of all the scattered waves which come from a given scattering point and fall on the antenna, only those inside a Fresnel zone contribute to the signal at the antenna output. This angle is inversely proportional to the square root of the range and is about equal to the beam width of the antenna at the limit of the near field, which is about 120 km for the Arecibo antenna at 430 MHz. This means that at stratospheric heights the effective angle is about 2.5 to 2.8 wider than the far-field beam width. A scatter moving at 10–20 m s^{-1} transverse to the antenna beam would have maximum projected velocities into the antenna on the order of 0.017–0.034 m s^{-1} in the far field, and 0.043–0.095 m s^{-1} at stratospheric heights. Therefore, the expected spectrum broadening due to finite beam widths can be neglected when compared to the measured spectral width of 0.3 m s^{-1} and higher. The effect is even less than the above figures indicate, when one considers that only a few points, at the edge of a Fresnel zone, contribute with above-maximum Doppler shifts, and that it is the square of the ratio of these velocities that determines the importance of finite beamwidth relative to turbulence broadening.

Other causes of spectral broadening are shear and antenna motion. Shear broadening will be discussed later, and there is no antenna motion broadening, since the antenna is kept motionless during every integration.

The existence of a strong angular dependence in the received echo power has been found by VHF radars, and is discussed in terms of partial specular reflection from the zenith direction (see Balsley and Gage, 1980). To examine the angular dependence of the observed parameters in the UHF range in more detail, we conducted a short experiment, changing the zenith angle between zero and 15°. Fig. 4 shows the echo power, the spectral width, and the line-of-sight velocity versus the zenith angle. Each point is an average over 3 min and the 14–19 km height range, in which the zonal velocity decreases from 30 to 10 m s^{-1}. The zenith angle was changed from 15° toward zero, then back to 12.5°. The antenna did not move while the integration was taking place. Open circles show the data taken on the way toward zenith, and black circles show the other. This experiment took about 70 mins. The data in the zenith direction are not used, because the stratospheric echo, having almost no radial velocity component, cannot be distinguished from the strong ground clutter contribution. The data at 1° zenith angle, although plotted in the figure, should also be treated with caution. The echo power is compensated for the dependence of antenna efficiency on zenith angle.

The echo power shows no appreciable angular dependence between 2.5 and 15°. Specular reflection, if it exists, would have to be confined to angles less than 2.5°.

The spectral width shows an interesting linear angular dependence. It can be explained most easily by the shear broadening effect. The difference in Doppler shift of the upper and lower boundary of the turbulent layer moving with the vertically sheared background flow contributes to a broadening of the echo power spectrum when the antenna is pointing obliquely. This broadening is proportional to the sine of the zenith angle, which is almost linear within 15°. If we assume this angular dependence is caused by the shear broadening effect alone, we can then estimate the zonal velocity difference over the turbulent layer to be ~1.2 m s^{-1} from the gradient of the fitted straight line. If we take the meridional velocity difference over the layer, which is not observed, to be the same as that of the zonal difference, we can estimate the vector velocity difference over the layer to be about 1.7 m s^{-1}. If we postulate that the local background shear is critical (Richardson number ~0.25) around the turbulence layer, it becomes about 40 m s^{-1} km^{-1} around the tropopause height. We can thus estimate the thickness of the layer to be about 40 m.

There is another way to estimate the thickness of the layer (Woodman and Guillen, 1974). By extrapolating the straight line toward zero zenith angle, we obtain the true spectral width of about 0.3 m s^{-1} in the line-of-sight velocity scale. This velocity can be interpreted as the random velocity of turbulence caused by the shearing instability. Assuming that all of the available energy supplied by the background shear goes to the kinetic energy, except for the loss due to the work done against buoyancy (one-fourth of the available energy for a Richardson number of 0.25), we can evaluate the random velocity to be about one-seventh of the velocity difference over the layer under the critical shear condition. Thus the vector velocity difference over the layer can be estimated as 2.1 m s^{-1}. This value is in good agreement with the former estimate of 1.7 m s^{-1}, and leads to an estimate of the thickness of the layer of about 50 m. These estimates are also consistent with the fact that the thickness is usually less than 150 m, as mentioned before.

A good linear zenith angle dependence of the line-of-sight velocity supports the assumption used in determining the horizontal velocity that the vertical velocity is very small on the average.

4. An estimation of the turbulent eddy diffusivity

The eddy diffusivity due to stratospheric turbulence has been estimated using aircraft (Lilly *et al.*, 1974), balloon (Cadet, 1977), rocket trail (Rosenberg and Dewan, 1975) and coherent radar observations. With radar, the estimates are derived from echo power with the aid of balloon temperature data (Gage *et al.*, 1980), and from occurrence frequency and measured thickness of the turbulent layers (Woodman *et al.*, 1980). These different estimates resulted in a fairly wide range of values and different conclusions on the importance of stratospheric turbulence on vertical transport phenomena. We will examine here an alternative way, based on the observed width of the power spectra. A similar attempt was made for mesospheric observations (Cunnold, 1975), interpreting the spectral width in terms of the eddy diffusivity.

The three-dimensional wavenumber spectrum of the kinetic energy of turbulence in the inertial subrange is given by

$$S(k) = a\epsilon^{2/3}k^{-5/3}, \quad (1)$$

where a is the Kolmogoroff constant (~ 1.6), and ϵ is the energy dissipation rate. $S(k)$ integrated from k_0 to k_i is the kinetic energy of the turbulence per unit mass measurable by a radar, where k_i is the radar Bragg wavenumber and k_0 is the wavenumber associated with the largest vortex in a turbulent layer. Since k_i is 18 rad s^{-1} for the radar frequency of 430 MHz, and k_0 is 0.1 to 1 rad s^{-1} for the thickness of the layer discussed before, then $k_i \gg k_0$. The width of the echo power spectra (expressed in velocity units) can be interpreted as the representative velocity of turbulence, if the shear broadening effect is removed from the spectra. We may then write the energy per unit mass as

$$\int_{k_0}^{k} S(k)dk = \tfrac{3}{2}\sigma^2,$$

and thus we obtain

$$\epsilon = a^{-3/2}\sigma^3/(k_0^{-2/3}-k_i^{-2/3})^{3/2} \simeq a^{-3/2}\sigma^3 k_0\,, \quad (2)$$

where σ is the one-side $1/e$ echo power spectral width, which is 0.6 times the usual half-power spectral width. We should note that spectral parameters in the radar observations are mean values over the volume determined by the antenna beam width and the altitude resolution weighted by the echo power. The true average in a height sample is obtained by multiplying (2) by the fraction of the turbulence in an altitude resolution. We may use the ratio of the size of the largest vortex in a turbulent layer and the altitude resolution as a crude estimate of this factor, when the echo power of the altitude exceeds the detectable threshold. We simply put it zero when the echo power is below the threshold. Thus we obtain

$$\bar{\epsilon} \approx (2\pi/k_0)/\Delta h\epsilon = 2\pi a^{-3/2}\sigma^3\Delta h, \quad (3)$$

where $\bar{\epsilon}$ denotes the mean energy dissipation rate in an altitude resolution Δh. Eq. (3) shows that the energy dissipation rate is directly connected to the spectral width. The eddy thermal diffusivity is given by (Lilly *et al.*, 1974)

$$K_h = \bar{\epsilon}/3N^2, \quad (4)$$

where N is the Brunt-Väisälä frequency.

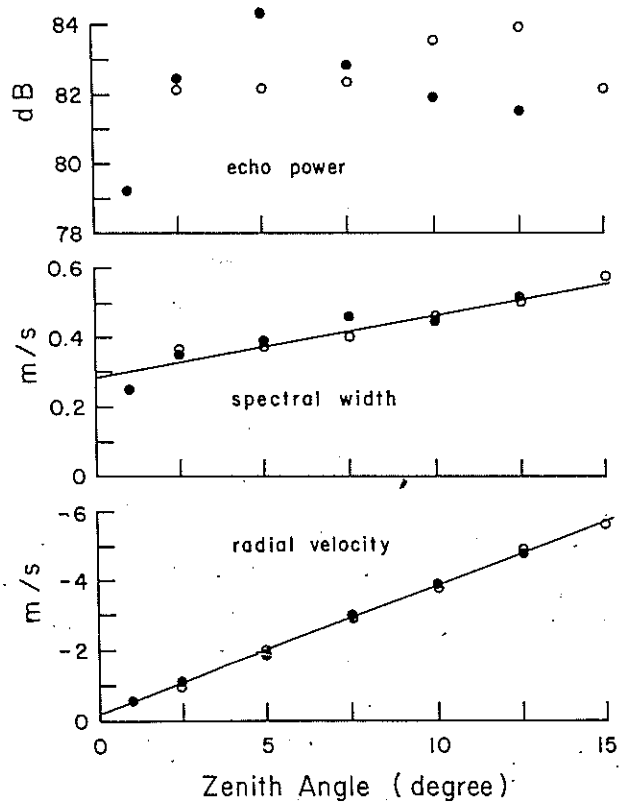

FIG. 4. The echo power, spectral width and radial velocity versus zenith angle. The zenith angle was changed from 15° to zero (open circles), then back to 12.5° (solid circles). All the measurements were done within about 70 min.

We are implicitely assuming that there is only one layer in a 150 m range gate. It is difficult to accept more than two, since they have an estimated thickness of 50 m. Even two is improbable, considering the structure of Fig. 1, which as we discussed earlier, shows layer separations on the order of a few hundred meters and more.

Fig. 5 shows one-hour mean profiles of the energy dissipation rate, Brunt-Väisälä frequency, and the eddy thermal diffusivity versus height. Since the Brunt-Väisälä frequency was obtained from seven rawinsonde measurements and interpolated to the day of the radar observation, it should be interpreted as a model. The variance in a month is, however, within 50% of the interpolated value. The resulting eddy diffusivity shows a clear maximum of about 1 $m^2\ s^{-1}$ around 16 km, the tropopause height. The average eddy diffusivities in the upper troposphere and the lower stratosphere are ~ 0.3 and 0.1 $m^2\ s^{-1}$, respectively, with comparable energy dissipation rates of $\sim 2 \times 10^{-4}$ $m^2\ s^{-3}$. The range of values of the energy dissipation rate, and its peak at the tropopause height, are in a general agreement with the higher-latitude Sunset radar observations (Gage *et al.*, 1980). The eddy diffusivity is about an order of magnitude larger than those reported in Lilly *et al.* (1974), but

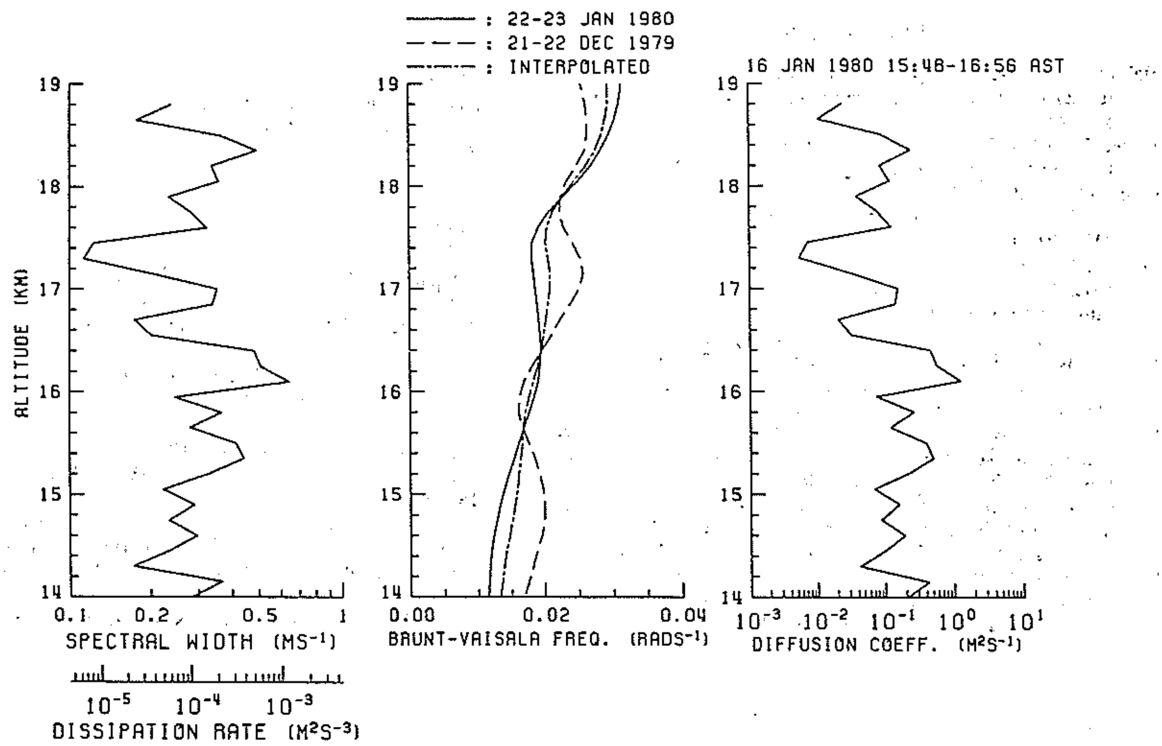

FIG. 5. One-hour mean profiles of the energy dissipation rate, the Brunt-Väisälä frequency, and the eddy thermal diffusivity.

agrees well with the mean value given by Woodman *et al.* (1981). Values of this order are sufficient to explain the residence times of radioactive debris and density profiles of tracer constituents in the lower stratosphere. Hence, they also show the importance of stratospheric turbulence in vertical transport in the lower stratosphere.

The advantage of this method of estimating the turbulence parameters is that instantaneous values of the energy dissipation rate can be deduced from the radar data alone. Though the eddy diffusivity depends on temperature through the Brunt-Väisälä frequency, it seems sufficiently stable to allow interpolation between balloon observations. The disadvantage may be that a very accurate value of the spectral width is required, since the parameters depend on its cube. The assumption used in determining the fraction of the turbulence in a height sample may add an error of a factor of 2 or so.

5. Summary

Some interesting results related to stratospheric turbulent layers, based on a high-altitude resolution experiment using the Arecibo 430 MHz radar, were discussed.

The turbulent layers appear as many discrete layers separated by a few to several hundred meters. The thickness of the layers is usually less than the altitude resolution of 150 m, and estimated to be about 50 m on the average, from a zenith-swinging experiment. Descending layers are observed in the troposphere. Layer altitude is more stable in the stratosphere. The horizontal extent of a layer in the lower stratosphere is estimated to be on the order of 100 km.

Observed echo power shows a good correlation with the vertical shear of the horizontal velocity, and with the spectral width of the echo, as is easily understood by an intuitive model.

The turbulent energy dissipation rate and the eddy thermal diffusivity were evaluated based on the observed width of the echo power spectra. A good agreement was found with other radar investigations of these parameters. Values on the order of 0.2 $m^2\ s^{-1}$ are found for the eddy diffusion coefficient in the lower stratosphere, showing the importance of the stratospheric turbulence on vertical transport phenomena.

Acknowledgments. The Arecibo Observatory is part of the National Astronomy and Ionosphere Center, which is operated by Cornell University under contract with the National Science Foundation. This work was supported partially by the National Aeronautics and Space Administration under NASA Order No. W-14,569.

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