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# ALTITUDE SMEARING DUE TO INSTRUMENTAL RESOLUTION IN MST RADAR MEASUREMENTS

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Abstract. Altitude smearing involved in MST radar measurements with low altitude resolutions is discussed, using the data obtained at Arecibo with an altitude resolution as high as 150 m. The smearing is of great concern in the case where the altitude resolution is as low as 3 km. The wind velocity deduced with the 3-km resolution is biased from the true velocity where the wind shear is large. The bias of the order of 10 m/s persists for hours. Moreover, there appear spurious velocity fluctuations, which do not correspond to any of the fluctuating components in true velocity variation. Also, most of detailed layered structures detected with 150-m altitude resolution are smeared out with 3-km resolution. The present result strongly suggests that we should be careful enough in interpreting MST radar data obtained with altitude resolutions as low as 3 km.

## 1. Introduction

During the past few years it has been generally realized that the MST (mesosphere-stratosphere-troposphere) radar technique promised a new efficient way in understanding the dynamical process occuring in the middle atmosphere (e.g., Balsley and Gage, 1980). This technique, which incorporates sensitive coherent VHF and UHF radars, is capable of measuring wind velocity averaged over a volume illuminated by a radar pulse.

The vertical extent of the illuminated volume, which gives an effective altitude resolution in measurement, is determined by the instrumental response of the radar system. In the case where the radar pulses are uniformly scattered everywhere within the volume, the volume-averaged velocity is precisely representative of the true mean wind of the volume. However, the scattering generally occurs from discrete layers of the turbulence, and the layers in the middle atmosphere are usually separated by more than 1 km (Woodman, 1980). This would cause smearing of the wind profile when measured by a radar with altitude resolutions lower than 1 km, since the velocity measured can be highly dependent on a small portion of the volume.

Recent technical development at Arecibo (18.3°N, 66.8°W) has attained an altitude resolution as high as 150 m in the lower stratospheric measurements with the 430-MHz radar (Woodman, 1980). However, the resolution at Jicamarca (12.0°S, 76.9°W) has been 3 km, or 20 times that of Arecibo, since the first middle atmospheric measurement performed there in 1971 (Woodman and Guillén, 1974). Also, the values are considerably larger than 150 m, normally 1 to 2 km, for most of other working radars. Therefore, it seems quite timely to assess now how representative the wind velocities measured with lower instrumental resolutions are. It was suggested by Woodman (1980) that the high-resolution data of the Arecibo radar can be used for this purpose. In the following, the altitude smearing of wind velocity along with echo power and spectral width will be discussed.

## 2. Method and model computation

First of all, we assume that echoes coming from turbulence layers within a volume illuminated by a radar pulse add up incoherently. Although this may not be valid for specularly reflected echoes of vertically incident waves, it is a good approximation for turbulent scattering of obliquely incident waves. Secondly, we consider the case where parameters such as the echo power, mean Doppler velocity or wind velocity, and spectral width are estimated from the zeroth, first, and second order integral moments of the echo power spectra, respectively. This is the simplest and most commonly used way of evaluating spectral parameters.

Finally, we assume that the overall instrumental response of a radar system is given by a raised cosine function of the following form;

$$W(h) = \begin{cases} \dfrac{1+\cos(\pi \frac{h}{\Delta h})}{2} & (- h \le h \le \Delta h) \\ 0 & (h < - h,\ h > \Delta h) \end{cases} \quad (1)$$

where h is the altitude, and Δh is the half power response which gives an effective altitude resolution in measurements with this radar system. No substantial change will be invoked in the following discussions, even if more plausible expressions, more faithful to the transmission/reception characteristics than Eq. (1), are employed.

In usual MST radar measurements, more than one turbulent layers seem to coexist within Δh so that the echo power spectra of individual layers cannot be separately observed as they are, but rather they are convolved with W(h) when observed. The smeared power spectra is given as

$$S_s(\omega,z) = \frac{\int W(h-z)\cdot S(\omega,h)\, dh}{\int W(h-z)\, dh} \quad (2)$$

where $S(\omega,z)$ is the unsmeared echo power spectrum at an altitude z. Radar deduced spectral parameters at this altitude are obtained by taking integral moments of $S_s(\omega,z)$. Spectral parameters evaluated from $S(\omega,z)$ and $S_s(\omega,z)$ are related by

$$P_s(z) = \frac{\int W(h-z)\cdot P(h)\, dh}{\int W(h-z)\, dh} \quad (3)$$



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$$V_s(z) = \frac{\int W(h-z)\cdot V(h)\cdot P(h)\ dh}{\int W(h-z)\cdot P(h)\ dh} \tag{4}$$

$$\sigma_s^2(z) = \frac{\int W(h-z)\{\sigma^2(h)+V^2(h)\}\cdot P(h)\ dh}{\int W(h-z)\cdot P(h)\ dh} - \left\{\frac{\int W(h-z)\cdot V(h)\cdot P(h)\ dh}{\int W(h-z)\cdot P(h)\ dh}\right\}^2 \tag{5}$$

where P, V, and $\sigma$ are echo power, mean Doppler velocity and spectral width obtained from $S(\omega,z)$, respectively, and the suffix s indicates the radar-deduced value. It is worth mentioning that $V_s$ is affected by the altitude profile of P, while $\sigma_s$ by those of P and V.

A simplified model is provided in Fig. 1 in order to illustrate how the spectral parameters are smeared. The solid line in the left, middle and right portions of this figure show the profiles for P, V and $\sigma$, respectively. Each scale is in an arbitrary unit. In this model, two scattering layers with a unit half-power width are assumed to exist among weak background scatterers. The relative echo power of both layers is 100 while that of the background is unity. It is also assumed in this model that the wind velocity increases linearly with altitude, and that the spectral width remains unity throughout the altitude range considered.

Broken lines are smeared profiles deduced for the case where $\Delta h$=20. When $\sigma_s$ is calculated, the radar beam is assumed to be tilted from the zenith by an angle of 10°. Significant smearing is observed in all of the three spectral parameters by the presence of strong scattering layers in the range where horizontal winds are sheared with altitude. The velocities both above the upper and below the lower scattering layers are fixed to the values at the layers' altitudes. The deduced spectral width is broadened by the effect of the wind shear. The broadening becomes larger as the radar beam is tilted more.

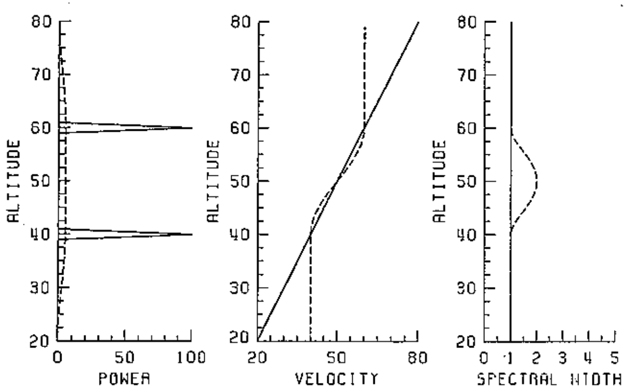


Fig. 1 Altitude profiles of echo power (left), wind velocity (middle) and spectral width (right). Each scale is given in an arbitrary unit. The solid lines are the models employed, while the broken lines are the smeared values observed by a radar with an effective altitude resolution of 20.

3. Computation based on observational data

In the following, some of the 150-m altitude resolution data obtained recently by the Arecibo radar (Sato and Woodman, 1981) are employed as a model representing real scattering properties of the lower stratosphere. The value of 150 m is not good enough to investigate detailed interior structure of individual turbulence layers, but it seems sufficient to identify intense turbulence layers from each other. Similarly to the preceding section, it will be shown here how the spectral parameters are smeared when measured with the altitude resolution 20 times that of the model.

The 150-m altitude resolution data are shown by the solid lines in Fig. 2. Each spectral parameter was estimated every 1 minute by fitting a Gaussian function to the measured power spextra in a least square sense (Sato and Woodman, 1980). The antenna beam was tilted eastward from the zenith by an angle of 10° during the observation, and the velocity is scaled to give the zonal component of the horizontal wind velocity. The layer observed around 16 km in the power profile is the intense turbulence layer usually observed near the tropopause over Arecibo. Also marked in this data is the intense vertical shear accompanied with the zonal wind around the tropopause.

The broken lines show smeared profiles. With the 3-km altitude resolution, it seems impossible to distinguish the detailed layered structure detected with the 150-m altitude resolution. A difference as large as 10 m/s is observed between the model and the smeared velocities around the altitude of 14-15 km. This velocity offset is caused by the intense turbulence layer existing in the sheared-wind region. It should be noted that the offset cannot be accounted for by simply averaging the model velocity along the altitude without weighting with echo power. In the case where the wind shear is small, on the other hand, the smearing is not so significant, even if the echo power is locally large as obsereved below 13 km. The spectra are generally broadened by the smearing, and is principally determined by the broadening effect caused by velocity difference in a 3-km vertical extent, where the wind shear is large.

Fig. 3 shows temporal variations of the model and the smeared wind velocities at five altitudes around the tropopause. It should be noted that the two variations seem to be completely unrelated. The difference between the two variations persists for hours. The two variations generally do not coincide with each other even by averaging them respectively over a few hours. It is due to the fact that the intense turbulence layers, existing in the sheared-wind region, generally keep their dominancy for more than a few hours without changing both their altitudes and intensities considerably (Sato and Woodman, 1981). There also appear short-period fluctuations in the smeared velocity variations, which are generally found to be spurious, quite different from the ones found in the model velocity variation. The fluctuations are considerably modulated by short-period fluctuations of the echo power. Thus, Figs. 2 and 3 strongly suggest that we should be

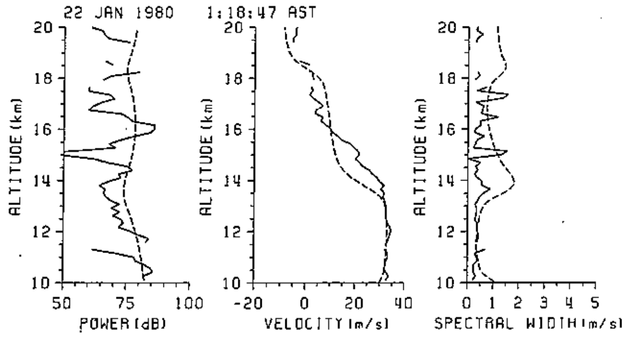


Fig. 2 Same as Fig. 1 except that parameters observed at Arecibo on 22 January 1980 are used for the model profiles. The echo power is relative value in decibels with an arbitrary reference level, while the mean Doppler velocity is scaled to the zonal wind velocity.

especially careful in interpreting velocity data taken with low altitude resolutions.

## 4. Discussions and concluding remarks

In an altitude extent of $\Delta h$, wind velocities vary due to vertical shear of $dV/dz$. The largest difference between the velocities at the center and on the top (or the bottom) of $\Delta h$, that could be conceived when the shear is constant within $\Delta h$, is $(dV/dz)\Delta h/2$ m/s. The largest possible wind shear can be estimated from the critical shear condition in which the Richardson number becomes 0.25. For an isothermal atmosphere of 200 K, this shear is about 0.04/s. Thus, it seems probable that at some altitude, a radar deduces a wind velocity which differs at most by $0.02\ \Delta h$ m/s from the true velocity at that altitude. This value corresponds to 3 and 60 m/s for 150-m and 3-km altitude resolutions, respectively. The 60-m/s velocity difference would result in spectral broadening as large as 10 m/s when measured in the oblique direction with 10° off from the zenith. However, the

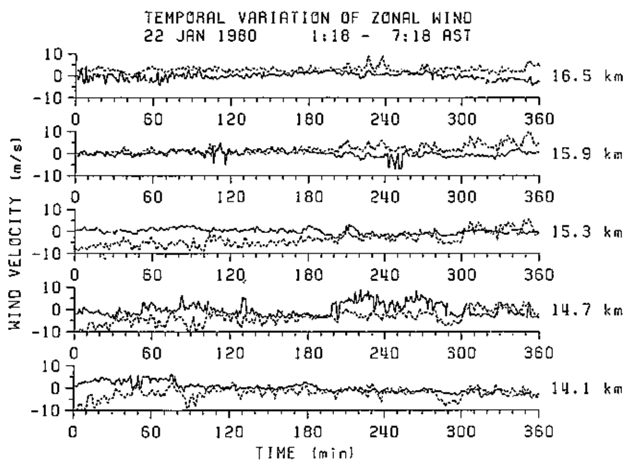


Fig. 3 Six-hour velocity variations at five altitudes around the tropopause. The model and the smeared velocities are given by solid and broken lines, respectively. The model velocities are the zonal wind velocities observed at Arecibo during 0118-0718 AST on 22 January 1981.

broadening deduced for the 3-km altitude resolution does not exceed 1.7 m/s as shown in Fig. 3. Therefore, it is expected that actually the velocity difference may be of the order of 1/6 of the value estimated for the extreme case.

In conclusion, it is conceivably possible that the altitude smearing is of great concern in MST radar measurements in the following two aspects, in the case where the altitude resolution is as low as 3 km. First, the velocity deduced with 3-km altitude resolution may possibly be biased from the true velocity in the case where the wind shear is large. The offset persists for hours, and can be as large as 10 m/s. There also appear spurious velocity fluctuations, which do not exist in true velocity variations. Second, all of the profiles of wind velocity, spectral width and echo power are considerably smeared in altitude when measured with 3-km altitude resolution. The detailed layered structure detected with 150-m altitude resolution cannot be distinguished with 3-km altitude resolution.

We have discussed here the effects of the altitude smearing in radar observations based on the tropospheric and stratospheric data. However, the conclusion obtained here is, in principle, applicable also to the mesospheric observations because we simply made use of the nature of multiple scattering layers within the altitude resolution moving at different velocities. Since layered scattering structures similar to those observed in the stratosphere are often found in the mesosphere by the high resolution SOUSY VHF radar (e.g., Czechowsky et al., 1979), it is quite likely that mesospheric observations made with a poor resolution system suffer from the smearing discussed here.

Lower altitude resolution of some of the radars in use, especially of the Jicamarca radar, is strongly hoped to be improved soon, in view of a possibility that the Jicamarca radar, the only radar situated in the tropics (12.0°S, 76.9°W), would promote much the study of various dynamical process occuring in the tropical middle atmosphere (e.g., Fukao et al., 1981).

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## References

Balsley, B. B., and K. S. Gage, The MST radar technique: Potential for middle atmospheric studies, Pure Appl. Geophys., 118, 452-493, 1980.

Czechowsky, P., R. Rüster, and G. Schmidt, Variations of mesospheric structures in different seasons, Geophys. Res. Lett., 6, 459-462, 1979.

Fukao, S., K. Aoki, K. Wakasugi, T. Tsuda, S. Kato, and D. A. Fleisch, Some further results on the lower stratospheric winds and waves over Jicamarca, J. Atmos. Terr. Phys., 43, 649-661, 1981.

Sato, T., and R. F. Woodman, Spectral parameter estimation of CAT radar echoes in the presence of fading clutter, Proc. 19th Conf. on Radar Meteorol., pp. 568-574, Amer. Meteorol. Soc., Boston, 1980.

Sato, T., and R. F. Woodman, High-altitude resolution observations of upper-tropospheric and lower-stratospheric winds and waves by the Arecibo 430-MHz radar, submitted to J. Atmos. Sci., 1981.

Woodman, R. F., High-altitude resolution stratospheric measurements with the Arecibo 430-MHz radar, Radio Sci., 15, 417-422, 1980.

Woodman, R. F., and A. Guillén, Radar observations of winds and turbulence in the stratosphere and mesosphere, J. Atmos. Sci., 31, 493-505, 1974.

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