ABSTRACT

Several differences in velocity profiles were observed in the high- and low-mode data collected by the 915-MHz wind profiler at San Cristóbal Island, Galápagos. Those differences are important because they prevent us from merging the data into one set. Some types of differences are expected because of the different volumes averaged by each mode. Others, however, are caused by violations of the assumptions behind the radar equation. The radar equation has two main assumptions that are violated: (1) the reflectivity is constant over the volume and (2) the range to the volume is large with respect to the pulse length. Johnston et al. (2002) proposed a more general radar equation without those assumptions.

This research is based on Coleman’s (2002) work, which assumes that the reflectivity gradient (RG) beyond 1.5 km was –2.17 dB/km, and on Burt’s (2003) results that found a linearly varying profile from 1.0 to 1.5 km. We test 3 different RG profiles in order to get an idea of how much impact the variations in the RG found by Burt have in a practical sense.

The heights assigned to the high-mode data were corrected during eight 10-day long cases that we selected based on the warm and cold season. The results show a slightly better agreement with low-mode winds from 1.0 to 1.5 km. This suggests that the method applied works but needs some additional corrections and further research to determine what new corrections are necessary to complete the goals.

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1. Introduction:

Radars are used to measure atmospheric parameters including rain, wind and turbulence. The Aeronomy Laboratory (AL) of the National Oceanic and Atmospheric Administration (NOAA) designed radars called wind profilers to measure wind speed and direction as a function of altitude. Wind profilers send out a beam of radio waves that reflect off turbulence in the atmosphere. The profilers observe these atmospheric conditions in a volume. They have one vertical and 2 or 4 oblique beams. Figure 1 is an example of a five-beam profiler. Ultra High Frequency (UHF) profiling provides a high resolution for lower tropospheric winds in lower altitudes that the Very High Frequency (VHF) profiler cannot reach.

The data used for this project is from a three-beam UHF system and is collected in two altitude modes: the high and the low modes. Radar pulses have different lengths, and the trade off is sensitivity vs. resolution. If we send a long pulse, we have high sensitivity and can observe to higher altitudes but with low resolution. In the short pulse we observe the opposite, low sensitivity and therefore low height coverage but with high resolution. With both modes Doppler shifts are used to calculate horizontal winds at many heights.

Some differences in velocity profiles were observed in the past in the high and low mode profiles. Hartten and Gage (2000) show seasonal mean high and low mode profiles that have similar features at different heights (Figure 2). These can be caused by the
different resolutions of volumes averaged used by each mode or by wrong height assignments due to differences in the gradients of reflectivity in the atmosphere (Johnston et al., 2001). The former differences are expected and acceptable because they are from sampling techniques. However, the reflectivity related differences should not occur because the radar is illuminating the same volume of the atmosphere in both operating modes. The objective of this research is to fix the errors in height by analyzing the radar equation assumptions and the reflectivity gradients in high-mode data. Then we can improve the quality of the data so the AL can merge the two modes in one data set and get a better understanding of the dynamics of the Pacific area.

2. Technical Approach:

a. Previous Work

Harten and Gage (2000) presented observations of lower tropospheric winds obtained over three years at the Galápagos Islands. The data showed disagreements between modes in the three years of seasonal-mean profiles. Johnston et al. (2002) made comparisons of data taken by collocated Doppler wind profilers using different pulse lengths. These showed that the velocity profiles obtained at Christmas Island by longer pulses are displaced in height from profiles measured with shorter pulses (Figure 3). Johnston et al. (2002) gave an explanation for these differences and proposed a correction
based on a more general form of the radar equation. The radar equation has two main assumptions that are violated: (1) the reflectivity is constant over the volume and (2) the range to the volume is large with respect to the pulse length. Figure 4 is a schematic of a profiler beam, only focusing on one volume. On the left side is the standard common assumption which is that the $RG = 0$. Because all locations in the volume are equally reflective, the wind information returned from the volume is assigned to the center of the volume. On the other hand, in the right side of that figure is an illustration of the Johnston et al. assumption, which is that the $RG \neq 0$. Because either the top or bottom of the volume is more reflective, the wind information returned is “weighted” towards the more reflective altitude, and should be assigned above or below the center of the volume.

Figure 3: Long-term mean profiles of wind speed measured at Christmas Island. The profiles are mean values for Feb 1994, 1995, and 1996. (Johnston et al. 2001)

Figure 4: Schematic of one volume in a vertically pointing profiler beam, showing the standard common assumption and Johnston et al. assumption of reflectivity gradient ($RG$).
Coleman (2002) explored the effects of this more general radar equation and used a very simple profile of atmospheric reflectivity gradient, \( RG = -2.17 \text{ dB/km} \), to correct height assignments of some wind profiler data. He found that in applying this equation to the calculations, the number of discrepancies in the two modes was reduced. Burt (2003) worked with the reflectivity gradients in the lower troposphere based on wind profiler data at Galápagos and Tarawa, Kiribati. She compiled a climatology of the reflectivity gradients to better understand their vertical structures and variability. She found fairly constant RG aloft, with linearly decreasing RG from \( \sim 1.5 \text{ km} \) down to \( \sim 1.0 \text{ km} \). She also concluded that there was significant variability in the mean reflectivity gradient for the meteorological seasons of December, January and February (DJF) versus March, April and May (MAM).

b. Data

In this research some height corrections are made, using data collected by the Galápagos profiler from 1994 to 1999. The Galápagos Islands are located in the Pacific Ocean (Figure 5), about 600 miles west of Ecuador. A 915-MHz wind profiler with pulse lengths of 100 and 500 meters collected the data set. There are two modes in which the AL tropical 915-MHz (VHF) profiler collects data continuously: high and low altitude modes. The low mode gives very high-resolution data because it ranges over a limited altitude. The high mode gives coarser resolution data over a deeper layer of the lower troposphere. The individual profiles of velocity vectors are converted into half-hourly profiles before being used for dynamical research.
c. Methodology

In order to do the height correction, P. E. Johnston (2004, personal communication) made height correction tables for our analysis based on the general radar equation used by Coleman (2002). For every 1-meter in range, $r_0$, from 1000m to 8000m, equation (7) in Johnston et. al (2002) was solved iteratively to obtain the true peak of the integrand, $r_1$. In solving the equation, he used the parameters for the Galápagos high mode: $B_0\tau$ product $= 0.63$ and the pulse length $= 500$ m.
Equation (7) requires values of RG (denoted “M” in Johnston et al., 2002) at all ranges. At ranges beyond 1.5 km we assumed the same RG that Coleman used, -2.17 dB/km. From 1.0 to 1.5km, we used a linearly varying profile based on Burt's results. We tested three different RG profiles (Figure 5), in order to get an idea for how much impact the variations found by Burt have in a practical sense when corrections are done. Also we retested Coleman’s correction as a control of sorts. The 3 different non–constant correction profiles used were based on 19 monthly mean profiles of Tarawa RG plotted by Burt (2003). The values at range=1000m represent the mean, max, and min values obtained from those profiles. The profiles were from the following months:

<table>
<thead>
<tr>
<th>Year</th>
<th>Months</th>
</tr>
</thead>
<tbody>
<tr>
<td>1994</td>
<td>Oct</td>
</tr>
<tr>
<td>1995</td>
<td>Jan, Apr, Jul, Oct</td>
</tr>
<tr>
<td>1996</td>
<td>none</td>
</tr>
<tr>
<td>1997</td>
<td>Jan, Jul</td>
</tr>
<tr>
<td>1998</td>
<td>Jul, Oct</td>
</tr>
<tr>
<td>1999</td>
<td>Jan, Apr, Jul, Oct</td>
</tr>
<tr>
<td>2000</td>
<td>Jan, Apr, Jul, Oct</td>
</tr>
<tr>
<td>2001</td>
<td>Jan, Apr</td>
</tr>
</tbody>
</table>

Tarawa is in the western Pacific, in the warm pool, but Burt found the basic shape of the RG profiles seemed similar to those from Galápagos.

Figure 7 is the offset applied to the different correction tables. That offset is for negative values of RG. All our RG for that research are negative. If the RG <0 the reflectivity is greater in the low part of the volume and all shifts go down. That’s the reason that the weighting for the data assigned to our cases is from the lower part of the volume (see Figure 4). That figure also shows that close to the radar, the heights
corrections can be larger, and if the offset is far from the radar the corrections can be smaller. This shows the effect of what is called the “r-squared” term in equation (7) of Johnston et al., in other words the effect of no longer assuming that the range to the volume is much larger than the pulse length.

![Graph showing reflectivity gradient and offset](image)

Figure 6: Linearly varying profile of reflectivity gradient from 1 to 1.5 km

Figure 7: Offset of the peak of the integrand resulting from the RG profiles in Fig. 6.

We have picked eight 10-day long cases during which profiles were corrected (Table 1), and some of these cases are roughly the same as used by Coleman (2002). The cases were selected based on seasons: the warm season goes from February-May and the cold season goes from July-October. The atmospheric conditions in the selected cases were that the Sea Surface Temperature in the warm season is warmer than the normal in some cases. In the cold season some cases, like September 1995 and August 1999, are colder than the normal. In July 1997 and September 1997, the SST is warmer than usual.
This is due to the El Niño-Southern Oscillation (ENSO). In the colder cases we can observe the displacement features better. The reason of that is that the air over the ocean is cold and it cannot mix well in the atmosphere, and the features are better defined.

<table>
<thead>
<tr>
<th>Date</th>
<th>Niño 1+2 °C</th>
<th>Anomål ° C</th>
<th>Season</th>
<th>SOI</th>
</tr>
</thead>
<tbody>
<tr>
<td>May/10-19/1995</td>
<td>23.10</td>
<td>-0.57</td>
<td>warm</td>
<td>-0.27</td>
</tr>
<tr>
<td>Aug/10-19/1995</td>
<td>20.01</td>
<td>-0.77</td>
<td>cold</td>
<td>-0.48</td>
</tr>
<tr>
<td>Sept/10-19/1995</td>
<td>19.24</td>
<td>-1.2</td>
<td>cold</td>
<td>0.43</td>
</tr>
<tr>
<td>May/10-19/1997</td>
<td>26.71</td>
<td>2.47</td>
<td>warm</td>
<td>-1.48</td>
</tr>
<tr>
<td>Jul/10-19/1997</td>
<td>25.59</td>
<td>3.79</td>
<td>cold</td>
<td>-1.73</td>
</tr>
<tr>
<td>Sept/10-19/1997</td>
<td>24.40</td>
<td>3.96</td>
<td>cold</td>
<td>-1.55</td>
</tr>
<tr>
<td>Jul/10-19/1998</td>
<td>23.43</td>
<td>1.63</td>
<td>cold</td>
<td>0.88</td>
</tr>
<tr>
<td>Aug/10-19/1999</td>
<td>19.75</td>
<td>-1.04</td>
<td>cold</td>
<td>0.40</td>
</tr>
</tbody>
</table>

Table1: The 8 cases selected, the SST (°C) from Niño 1+2 and its anomal data, the seasons and the 5 Month Running Mean SOI.

First the high and low-mode data were interpolated to a uniform vertical grid. To do this we used an interpolator program created by C. H. Love and modified by L. M. Hartten that takes half-hourly profiler winds and interpolates them. The height offsets shown in Figure 7 were applied to the original high-mode data; the corrected profiles were interpolated to the same vertical grid. Mean profiles were then computed. We are not correcting the low mode data because the 100 m pulse height errors are considered quite small (Johnston et al, 2002).
3. Results and Discussion

Many plots were obtained from the analyzed 8 cases for the uncorrected and corrected data from Galápagos islands. A few cases are shown here, in figures 8-15. In the analysis of these plots we try to observe how the corrections work with the different RG profiles. These 10-day long cases reveal the structure of the wind speed and direction during those days.

a. Uncorrected Data

In most of the uncorrected original cases (Fig. 8-11) you can see the variations in height among the high and low mode profiles. In those plots you can see the differences in wind speed and direction. In the low mode the speed minimum occurs approximately in 1.1 km more or less in some of the cases, and in the high mode the minimum occurs in 1.4 km and up more or less in some cases. Those differences go from 1.0 to 1.5 km approximately in the eight cases. The May 1997 case (Fig 9) is an exception that does not show clear features. That case represents the warm season, and the sea surface temperature of this time was warmer than usual and for that reason this case doesn’t show clear features. The difference between both modes is approximately 300 m. Another feature observed in those plots is that the wind direction shifted somewhere between 1.0 and 1.5 km. In the lowest troposphere the winds are from the southeast and in the higher troposphere they are from the northeast.
Figure 8: May 1995 case, original wind speed and direction. That date represents the warm season. Blue lines represent the high mode, and red lines represent the low mode. In the wind direction the 0° represents north winds, 90° represents east winds, etc.

Figure 9: May 1997 case, original wind speed and direction. That date represents the warm season. Blue lines represent the high mode, and red lines represent the low mode.
Figure 10: July 1997 case, original wind speed and direction. That represents the cold season. Blue lines represent the high mode, and red lines represent the low mode.

Figure 11: September 1997 case, original wind speed and direction. That represents the cold season. Blue lines represent the high mode, and red lines represent the low mode.
b. Corrected Data

The corrected data (figures 12-15) show the effects obtained when we apply the correction to the original high mode data using the different reflectivity gradients in each case, based on Burt’s results. As you can see in Figure 11, results show slight changes in height in both wind speed and direction plots. Below 1.5 km we can see that the changes were bigger than above 1.5 km. The plots show that if the RG is more negative the lower part of the profiles are displaced further as shown in Figure 7. In Figure 12, which is the less interesting case because it does not have clear features, you can observe some corrections too. Not all shifted cases matched the original low mode. It is not the great change that we wished, but it is possible to observe that the method works. In some cases lower altitude data is overcorrected as you can see in Figure 11.

Figure 12: May 1995 case, corrected wind speed and direction. Blue dashed line, and red dashed lines represent the original high and low mode. The other solid lines represent the different RG corrected high mode data.
Figure 13: May 1997 case, corrected wind speed and direction. Blue dashed line, and red dashed lines represent the original high and low mode. The other solid lines represent the different RG corrected high mode data.

Figure 14: July 1997 case, corrected wind speed and direction. Blue dashed line, and red dashed lines represent the original high and low mode. The other solid lines represent the different RG corrected high mode data.
Figure 15: September 1997 case, corrected wind speed and direction. Blue dashed line, and red dashed lines represent the original high and low mode. The other solid lines represent the different RG corrected high mode data.

4. Conclusions

In this project we observed the main problem of the wind profiler. It has the same features at different heights when we created wind speed and direction profiles with the data. We tried to make some corrections to the high mode data from Galápagos islands during eight 10-days long cased that we selected. We used fairly simple reflectivity gradient profiles to do the corrections.

The corrections brought the high mode data closer to the low mode data, but did not made the biggest change that we hoped. In different cases the RG profiles worked better. The most obvious problem that we realized after the corrections was that the location of the features in the wind speed and direction profiles is approximately 1.5 km in the high mode data. Above that height, the corrections are all small, so do not help us
to bring the two modes into agreement. The results suggest that the applied method works, but needs some additional corrections in order to obtain better results. It shows us that only one or few RG profiles cannot be used to correct all half-hourly winds.

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References:


