A Surface Reference Technique for Airborne Doppler Radar Measurements in Hurricanes

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ABSTRACT

The operational algorithm for rainfall retrieval from the Tropical Rainfall Measuring Mission (TRMM) precipitation radar data requires a measurement of the path-integrated attenuation (PIA) as a constraint. This constraint is derived via the surface reference technique, whereby a measurement of the ocean surface in a raining area is compared with the ocean surface backscatter in a neighboring clear area. This method assumes that the surface backscatter difference is due only to the presence of rain, although variation in surface winds could also cause differences in the reference and rain measurements. An alternative surface reference method is to use a measurement of the surface winds and a backscatter model to predict the rain-free, or reference, cross section. Such an approach is developed here for airborne Doppler radar measurements in hurricanes. This approach provides an independent measurement of the reference backscatter, which is compared with the standard clear-air reference. The mean difference between the standard and Doppler-derived PIA is less than or equal to 1 dB; the rms difference is in the range 0.9–2.6 dB. In deriving the model function for backscatter estimation from wind measurements, the authors also find that the dependence of ocean backscatter on wind appears to saturate at high wind speeds at 25° incidence.

1. Introduction

Retrieval of the rain-rate profile from spaceborne radar measurements is complicated by the presence of attenuation at the higher frequencies used for these systems. The precipitation radar (PR) (Kozu et al. 2001) on the Tropical Rainfall Measuring Mission (TRMM) satellite operates at 13.8 GHz, where attenuation can be severe in heavy rain. In such cases the path-integrated attenuation (PIA) can be used as a constraint in rainfall retrieval, as is done in the operational algorithm for rainfall retrieval from the PR (Iguchi et al. 2000). The PIA is derived using the surface reference technique (SRT) (Meneghini et al. 2000), whereby a measurement of ocean surface backscatter in a raining area is compared with measurements in a neighboring clear area, and the difference between the two yields the PIA in the rain area. This method assumes that the ocean backscatter cross sections in the clear area and in the raining area are identical. Since ocean surface backscatter depends on surface winds, the SRT is also assuming that the winds in the clear and neighboring raining areas are nearly identical, which may not always be the case, as noted by Meneghini et al. (2000). Durden et al. (1995) used airborne radar and radiometer data to examine the SRT and found generally good agreement between the SRT method applied to the radar data and the PIA derived from the radiometer. There was essentially no bias, although the rms difference was on the order of 1 dB. Li et al. (2002) derived a wind retrieval method for the TRMM PR to estimate the effect of wind field variation on the surface reference technique. The mean difference between the wind field approach and the standard SRT was less than 1 dB, and the rms difference was 1–2 dB.

Further testing of the standard SRT could be accomplished if a direct measurement of the surface wind in the rain area were available. While this is not possible with the TRMM PR alone, it can be accomplished with an airborne Doppler radar in certain situations. Using the measured wind speed, the reference surface backscatter can be calculated using a backscatter model. This approach provides a reference that is independent of the standard SRT reference. The main assumption in this approach is that the backscatter is only a function of the surface wind; other effects, such as impact of rain drops on the ocean surface, are neglected. In this work we use the Doppler technique on airborne weather radar data in hurricanes, since their well-defined wind structure, in combination with Doppler measurements, can be used to provide an estimate of surface winds. We first discuss measurement of surface winds using airborne Doppler radar observations. Results are compared with independent measurements. Next, we discuss modeling of the ocean surface backscatter under high-wind
TABLE 1. ARMAR system characteristics.

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Operating frequency</td>
<td>13.8 GHz</td>
</tr>
<tr>
<td>Surface horizontal resolution</td>
<td>800 m</td>
</tr>
<tr>
<td>Range resolution (6 dB)</td>
<td>100 m</td>
</tr>
<tr>
<td>Range sampling</td>
<td>60 m</td>
</tr>
<tr>
<td>Antenna beamwidth</td>
<td>3.8°</td>
</tr>
<tr>
<td>Antenna scan angle</td>
<td>±25°</td>
</tr>
<tr>
<td>Swath at surface</td>
<td>10 km</td>
</tr>
<tr>
<td>Pulse repetition frequency</td>
<td>5 kHz</td>
</tr>
</tbody>
</table>

conditions. There are limited high-wind backscatter observations in the literature, so these results are significant apart from their application to measurement of the path-integrated attenuation in rain. Finally, we combine the surface wind measurement technique and the backscatter model to estimate the reference backscatter cross section and show comparisons of the standard SRT measurement versus our Doppler-derived measurement.

2. Doppler surface wind measurement

The Airborne Rain Mapping Radar (ARMAR) is a 13.8-GHz radar that operates on the National Aeronautics and Space Administration (NASA) DC-8 aircraft. It measures reflectivity, Doppler, and dual-polarization parameters. ARMAR’s antenna scans ±25° in the cross-track dimension. The aircraft operates at a nominal altitude of 11 km with a typical speed of 215 m s\(^{-1}\). Table 1 provides the ARMAR system parameters; further details can be found in Durden et al. (1994). We consider ARMAR measurements as shown in Fig. 1. The aircraft is moving directly toward or away from the hurricane’s eye, so that the cross-track scanning plane is approximately parallel to the hurricane’s tangential wind. The radial wind above the boundary layer is assumed very small. The cross-track scan angle \(\theta\) is the angle of the radar beam in the scan plane relative to nadir, with positive \(\theta\) to the right of nadir. The vector \(\mathbf{k} = \sin \theta \mathbf{y} + \cos \theta \mathbf{z}\) is a unit vector that points along the radar beam within the scan plane. The radar measures a mean Doppler speed that depends on the component of hydrometeor motion along the radar beam. The mean Doppler speed is the first moment of the Doppler spectrum, which includes contributions from the mean air motion (i.e., wind), turbulence, wind shear, raindrop terminal velocities, and platform motion (Doviak and Zrnic 1984; Hildebrand and Moore 1990; Heymsfield et al. 1996). The mean of the Doppler spectrum is the sum of the wind, raindrop, and platform means. To estimate the mean wind motion along the beam, we need to remove the components of platform motion and raindrop terminal velocity along the beam. The platform motion is removed using the measured Doppler from the ocean surface across a scan. This is an operational version of the technique described in Durden et al. (1999).

If we let \(\mathbf{v}_{\text{r}} = v_{\text{r}} \mathbf{z}\) be the reflectivity-weighted mean raindrop terminal fall velocity, then following correction for platform motion, the radar measured Doppler speed \(v_{\text{d}}\) is

\[
v_{\text{d}} = v_{\text{r}} + \mathbf{v}_{\text{f}} \cdot \mathbf{k} = v_{\text{r}} + v_{\text{z}} \cos \theta,
\]

where \(v_{\text{r}}\) is the mean air motion along the radar beam. To estimate \(v_{\text{z}}\) from the measured \(v_{\text{d}}\), we need to remove \(v_{\text{r}} \cos \theta\). A typical way of doing this is to use a \(v_{\text{r}}-Z\) relation to estimate \(v_{\text{r}}\) from the radar reflectivity \(Z\). Using the \(v_{\text{r}}-Z\) relation from Atlas et al. (1973),

\[
v_{\text{r}} = 2.6Z^{0.107},
\]

we can estimate \(v_{\text{r}} \cos \theta\) for each range bin and scan angle \(\theta\), and estimate \(v_{\text{r}}\) from \(v_{\text{d}}\). This procedure is standard in Doppler weather radar analyses (e.g., Doviak and Zrnic 1984; Black et al. 1996; Lee et al. 1999). Terminal velocity correction and motion correction are both applied to data from the dual-beam NASA ER-2 Doppler (EDOP) radar (Heymsfield et al. 1996), which looks at nadir and 33° forward along the flight track. EDOP data are processed to retrieve winds in the along-track plane rather than the cross-track plane in the case of ARMAR. Prior to correction of the ARMAR data for platform motion and raindrop terminal velocity, the measured Doppler speeds are dealiased.

Following the corrections just described, we have the measured component of air motion \(v_{\text{a}}(r, \theta)\) along each radar beam at each range bin. We now wish to estimate the hurricane tangential wind from these measurements. Again, referring to Fig. 1, we have

\[
v_{\text{a}}(r, \theta) = \mathbf{v}_{\text{f}} \cdot \mathbf{k} = v_{\text{a}}(r, \theta) \sin \theta + v_{\text{z}}(r, \theta) \cos \theta,
\]

where \(\mathbf{v}_{\text{f}}\) is the wind vector with tangential and vertical components \(v_{\text{a}}\) and \(v_{\text{z}}\), respectively. We would like to recover these components at all \(r\) and \(\theta\) from the measured \(v_{\text{a}}(r, \theta)\). To use single-Doppler measurements to obtain both velocity components in the cross-track plane, we assume that the tangential and vertical wind components at a given altitude are constant for all scan angles. In other words, \(v_{\text{a}}\) and \(v_{\text{z}}\) are assumed to be functions of altitude only, not of \(\theta\) over the ARMAR swath, which is approximately 10 km wide at the surface. Assuming that data at \(N\) scan angles are acquired, we have \(N\) measurements of the wind from different...
angles. Since we are trying to estimate the two velocity components in the cross-track scan plane, measurements from only two angles would suffice (Heymsfield et al. 1996). Having $N$ measurements allows us to write the following overdetermined system of equations at a given altitude:

$$
\begin{bmatrix}
\sin \theta_1 & \cos \theta_1 \\
\vdots & \vdots \\
\sin \theta_n & \cos \theta_n
\end{bmatrix}
\begin{bmatrix}
v_y(\theta_1) \\
v_y(\theta_2) \\
\vdots \\
v_y(\theta_n)
\end{bmatrix}
= 
\begin{bmatrix}
v_y, \\
v_y, \\
\vdots \\
v_y,
\end{bmatrix}
$$

where $v_y(\theta_i)$ is the measured $v_y$ at angle $\theta_i$, after correction for platform motion and raindrop terminal velocities. This system can also be written in matrix notation:

$$
A \vec{v} = \vec{v}_m,
$$

where $A$ is the $N \times 2$ matrix of sines and cosines in (4), $\vec{v}$ is the wind vector $v_x, v_y, v_z$, and $\vec{v}_m$ is the vector of measured $v_y$ at each scan angle $\theta_i$.

Assuming that the measurement error is not a function of $\theta_i$, (4) or (5) can be easily solved via least squares (Strang 1973). The least squares estimate of $\vec{v}$ is found by solving the normal equations corresponding to (4) or (5). In matrix notation these are $A^t A \vec{v} = A^t \vec{v}_m$, where $A^t$ denotes the transpose of $A$. In component form we have

$$
\begin{bmatrix}
\sum \sin^2 \theta_1 & \sum \cos \theta_1 \sin \theta_1 \\
\sum \cos \theta_1 \sin \theta_1 & \sum \cos^2 \theta_1 \sin \theta_1 \\
\sum \cos \theta_1 \sin \theta_1 & \sum \cos^2 \theta_1 \sin \theta_1 \\
\sum \cos \theta_1 \sin \theta_1 & \sum \cos^2 \theta_1 \sin \theta_1
\end{bmatrix}
\begin{bmatrix}
v_y \\
v_y
\end{bmatrix}
= 
\begin{bmatrix}
\sum v_y \sin \theta_1 \\
\sum v_y \cos \theta_1
\end{bmatrix}.
$$

This is a $2 \times 2$ system of equations and can be solved analytically for $v_x$ and $v_y$. A separate set of normal equations can be written for the measurements at each altitude. Their solutions provide the tangential and vertical wind components at each altitude below the aircraft, with the assumption of constant wind versus scan angle.

Because clutter from the surface can contaminate near-surface radar return, it is not feasible to use the above-mentioned method to directly measure surface winds. Instead, we must translate the horizontal wind $v_x$ at 400-m altitude, derived by the Doppler method, to surface wind speed. Powell (1980) has evaluated several boundary layer models. He found that the simple approach of multiplying the wind at a few hundred meters altitude by 0.8 works well, coming in second in a comparison of five models. We use this approach here because of its simplicity and acceptable accuracy. This approach is also used by Young (1993) and Yueh et al. (2001).

To test this approach of retrieving surface winds, we compared with winds from two other sources. One source is the National Oceanic and Atmospheric Administration (NOAA) Hurricane Research Division (HRD) analysis of flight-level wind measurements. These surface wind analyses use a boundary layer model and an objective analysis package described by Powell and Houston (1996). A second source is the hurricane wind field model of Holland (1980). This simple model generates a wind field given the hurricane location, forward motion speed and direction, central and ambient pressure, and radius of maximum wind. This model was used by Young (1993) for comparison with spaceborne altimeter measurements and by Yueh et al. (2001) for comparison with spaceborne scatterometer measurements. The ARMAR data used here were acquired during the Third Convection and Moisture Experiment (CAMEX-3) in 1998. Examples are shown in Fig. 2.

Differences between the HRD and ARMAR winds may be caused by differences in the time at which the measurements were made. Data used in the HRD output are typically acquired over several hours, whereas the ARMAR data are acquired during a flight line that may last 10–15 min. In the case shown in Fig. 2a, Hurricane Bonnie had a forward motion of 5 m s$^{-1}$ and would have moved at least 50 km during the HRD analysis period. Also, it should be noted that in the interval 90–120 km, the ARMAR data had very low signal-to-noise ratio due to very low reflectivities; hence, Doppler data...
were not available. Each ARMAR data point was acquired during a single 1.5-s scan. While the output from Holland’s (1980) model for Hurricane Danielle in Fig. 2b is based on input from a much smaller time interval than the HRD analyses, the model itself assumes a symmetric storm circulation. Asymmetry in the wind field comes only from the storm motion, whereas real storms may have much greater asymmetry. In the case shown here, the agreement is reasonable, although the ARMAR winds have about 3 m s\(^{-1}\) rms fluctuations about the model winds. This could be Doppler measurement errors or real fluctuations in the wind speed not predicted by a simple model. As with Fig. 2a, regions with missing ARMAR data had reflectivities that were too low for Doppler retrievals.

The error in our wind retrieval approach depends on a number of factors, including the assumption of uniform winds within the ARMAR swath. Lack of wind uniformity may be particularly pronounced in highly convective areas, such as the eyewall region. A quantitative assessment of the correctness of the uniformity assumption is the magnitude of the rms error for the least squares solution. We generally find that the errors are 1 m s\(^{-1}\) or less, indicating that the uniform model used here is a good description of the wind field within the ARMAR swath. Another potential source of error is terminal velocity correction, which can be in error by up to 1 m s\(^{-1}\) (Heymsfield et al. 1996). However, our method is only weakly sensitive to terminal velocity correction. In fact, for a uniform reflectivity field, the terminal velocity correction has no impact on the retrieved \(v_t\). Sensitivity studies with our retrieval indicate that 1 m s\(^{-1}\) errors in \(v_t\) normally have a very small impact on the retrieved \(v_t\), (<0.2 m s\(^{-1}\)) and can be ignored relative to other error sources. We estimate the total error in the Doppler method to be approximately 3 m s\(^{-1}\), based on 1 m s\(^{-1}\) accuracy for velocity estimation and platform corrections (Durden et al. 1999), 1 m s\(^{-1}\) for errors due to deviations of the ARMAR scan plane from the tangential wind direction (assumed less than 15\(^\circ\)), 2.5 m s\(^{-1}\) accuracy for the simplified boundary layer model (Powell 1980), and 1 m s\(^{-1}\) for the uniform wind assumption. These error sources are assumed independent, so the total error is the square root of the sum of squared errors.

### 3. Backscatter modeling

Ocean backscatter is typically described theoretically using a two-scale model, in which the backscatter cross section per unit area \(\sigma^\circ\) is the sum of geometrical optics and Bragg scattering components (Durden and Vesecky 1985). While it is possible in principle to retrieve the wind speed using a theoretical \(\sigma^\circ\) model, in practice it is more straightforward and potentially more accurate to use an empirically derived model, a so-called model function. It has been found that \(\sigma^\circ\) versus wind speed \(U\) obeys a power-law relationship \(\sigma^\circ = \alpha U^\gamma\), and most model functions provide the coefficients \(\alpha\) and \(\gamma\) versus the scan, or incidence, angle. Schroeder et al. (1985) provide coefficients for airborne scatterometer measurements. Seasat spaceborne scatterometer measurements, used to derive the Seasat-A Satellite Scatterometer (SASS-2) model function, are reported in Wentz et al. (1984). More recent model function include the NASA scatterometer (NSCAT-1) model function (Wentz and Smith 1999) and its successor, NSCAT-2. Most of the data used to develop model functions were acquired at wind speeds less than 20 m s\(^{-1}\).

To develop a high-wind backscatter model at nadir and 25\(^\circ\) (the range of ARMAR scan angles), we used ARMAR measurements of the ocean in rain-free areas, combined with winds from the HRD analyses and with winds derived from ARMAR using the technique described in the previous section. ARMAR backscatter measurements have been routinely used to estimate ocean \(\sigma^\circ\) for calibration purposes (Durden et al. 1994, 1997). ARMAR \(\sigma^\circ\) is found by integrating reflectivity over surface range bins. This and other approaches for estimating \(\sigma^\circ\) are discussed by Caylor et al. (1997). Rain-free areas are defined as those areas having small attenuation. For the HRD wind cases, ARMAR \(\sigma^\circ\) measurements are classified as rain free if all ARMAR range bins had reflectivity less than 20 dBZ. For the ARMAR wind cases, this threshold had to be increased to 28 dBZ to have enough cases for analysis. These cases require a large enough reflectivity that the ARMAR Doppler is measurable but a small enough reflectivity that the surface \(\sigma^\circ\) is not biased by attenuation. To verify that the bias is small, we applied a stochastic rain profiling algorithm (Durden and Haddad 1998) to the rain profile for each \(\sigma^\circ\) measurement. The estimated two-way PIA for all the \(\sigma^\circ\) measurements is indeed small (<0.4 dB).

Figure 3 shows the ARMAR measurements of \(\sigma^\circ\) versus the wind speed \(U\) at nadir and 25\(^\circ\) incidence. A total of 1325 points is shown for nadir; of these, 1013 are HRD wind cases and 312 are ARMAR wind cases. For 25\(^\circ\) a total of 957 points is shown; of these, 705 are HRD wind cases and 252 are ARMAR wind cases. The data were acquired in Hurricane Bonnie on 23, 24, and 26 August 1998; in Hurricane Danielle on 30 August 1997; and in Hurricane Georges on 21 September 1998. Along with the nadir ARMAR data in Fig. 3a, we have plotted the SASS-2 model function, \(\sigma^\circ = 42.66U^{-0.58}\). There is good agreement between SASS-2 and both sets of ARMAR data at nadir. There is no apparent saturation of \(\sigma^\circ\) at high wind speed at normal incidence; \(\sigma^\circ\) is a weakly decreasing function of wind speed even at high winds, similar to the results of Young (1993), who used nadir-looking altimeter measurements.

For 25\(^\circ\) incidence (Fig. 3b), we have plotted the ARMAR data, along with the SASS-2 and NSCAT-2 upwind models. At low wind speeds both the models and the ARMAR data are fairly close. At high wind speeds both the SASS-2 and NSCAT-2 models overestimate \(\sigma^\circ\).
This model is valid for winds of 45 m s\(^{-1}\) or less. Yueh et al. (2000) present geometrical optics calculations of \(\sigma^o\) in high winds and find that \(\sigma^o\) saturates more quickly at smaller incidence angles. This may explain why our observations show saturation for wind speeds in the 20–30 m s\(^{-1}\) range, while observations at much larger angles, where Bragg scattering is dominant, indicate wind speed sensitivity for winds up to 40 m s\(^{-1}\) or more (Yueh et al. 2001).

In the next section we use the SASS-2 model at nadir and (7) at 25°, along with the Doppler-measured surface wind, to estimate the reference \(\sigma^o\). As can be seen in Fig. 3 there is considerable scatter in the measurements about the model functions. The difference between measurement and model function depends on angle and wind speed, with the rms difference at nadir being 0.9 dB at all wind speeds, 1.5 dB at 25° for wind speeds below 10 m s\(^{-1}\), and 0.7 dB at 25° for high wind speeds. The scatter in Fig. 3 is similar to scatter reported in other studies and is due to the inherent variability of \(\sigma^o\) at a given wind speed, as well as errors in the measured wind speed. In the previous section the wind speed error for the ARMAR Doppler technique was estimated to be roughly 3 m s\(^{-1}\). Although the errors in the HRD winds are not known, the scatter in Fig. 3 for both the open circles (HRD winds) and filled circles (ARMAR winds) are similar. Based on the slope of the model functions for wind speeds near 15 m s\(^{-1}\), a 3 m s\(^{-1}\) error in wind speed would result in an error in \(\sigma^o\) of around 0.6 dB. The \(\sigma^o\) error would be smaller at higher wind speeds. The scatter in Fig. 3 suggests an error of 1–2 dB in our estimated reference \(\sigma^o\), and we expect better performance at higher wind speeds due to the reduced sensitivity to wind speed error, particularly at 25° incidence.

### 4. Application to hurricane measurements

We now have a method for using the radar Doppler winds to estimate the surface winds in hurricanes and a method to estimate \(\sigma^o\) using those winds. Figure 4 shows results for a case from Hurricane Bonnie on 26 August 1998. Figure 4a shows results at nadir, while Fig. 4b shows results at 25° scan angle. Plotted in each case are the measured \(\sigma^o\) (attenuated by rain when present), the reference \(\sigma^o\) derived by the standard SRT, and the Doppler-based reference \(\sigma^o\). The two-way PIA can be derived by differencing the observed \(\sigma^o\) and the reference \(\sigma^o\). The standard SRT as implemented here is simplified for use with airborne data. Of the several types of references implemented for the TRMM PR (Meneghini et al. 2000), only the along-track spatial reference is implemented here. Specifically, the reference \(\sigma^o\) is the last available rain-free observation of the surface. When the algorithm encounters a surface backscatter in rain that is larger than the reference measurement, the PIA is set to zero and the reference is equal to the measurement for this case.

The areas of heavy rain in Fig. 4 can be seen by the
reduction in the measured $\sigma^0$ values. At nadir the standard SRT is somewhat larger than the Doppler reference $\sigma^0$. At 25° the standard SRT is lower than the Doppler estimate. Furthermore, the standard SRT tends to change rapidly in some areas, as new clear areas are encountered. In large areas of high attenuation, the standard SRT remains constant due to lack of new clear areas in which to update the reference $\sigma^0$. The Doppler SRT provides a smoother reference $\sigma^0$, since it follows wind speed changes in rain areas.

Table 2 shows the mean of the standard SRT $\sigma^0$ minus the Doppler SRT $\sigma^0$ at both nadir and 25° incidence, as a function of the Doppler SRT $\sigma^0$. All $\sigma^0$ are two way. The results in Table 2 are based on 14 ARMAR radial flight lines in Tropical Cyclone Oliver in 1993 near Australia (nadir only) and in Hurricanes Bonnie and Danielle in 1998 during CAMEX-3. There is no clear trend in the mean difference versus $\sigma^0$ at nadir; both negative and positive differences occur. At 25° incidence the difference is mostly negative but small, indicating that the Doppler $\sigma^0$ is usually larger than the standard SRT $\sigma^0$. This could be due to higher winds in the hurricane creating a larger $\sigma^0$ than found in a nonraining clear area with lower winds. In all cases the mean difference is no larger than 1 dB. Also shown in Table 2 are the rms differences, which range from 0.9 to 2.6 dB.

The observed differences in the two techniques are similar to the expected errors for the Doppler reference estimate. The lack of large differences between the two techniques is likely related to the lack of sensitivity of $\sigma^0$ to wind speed at high winds. If the differences here are interpreted as variability in the standard SRT, they are similar to the assumed $\sigma^0$ variability in the TRMM PR rainfall algorithm, designated 2A25, and described by Iguchi et al. (2000). Over the ocean, 2A25 assumes a $\sigma^0$ standard deviation of at least 1 dB, which is roughly consistent with our results. The results found here require some caution when applied to TRMM PR data, since the ARMAR horizontal resolution of 800 m may allow some small clear areas to be found even within areas of heavy rain. The TRMM PR resolution of 4.3 km could smooth such areas, causing them to be flagged as rain areas. Our results are similar to those in Durden et al. (1995) and especially Li et al. (2002) using different techniques.

5. Conclusions

The surface reference technique is normally implemented using clear-air surface measurements to derive a reference $\sigma^0$. This reference is compared with $\sigma^0$ measurements in rain to estimate the $\sigma^0$, assuming that the surface $\sigma^0$ is the same as in clear air. We have developed an alternative surface reference technique that can be used in hurricanes. With this technique Doppler radar measurements of surface winds are used to drive a $\sigma^0$ model. The winds are retrieved using a least squares approach with a simple correction for the boundary layer. The $\sigma^0$ model was derived from ARMAR measurements and wind measurements from both HRD and ARMAR. At 25° incidence ARMAR $\sigma^0$ measurements versus wind speed appear to saturate at wind speeds in the 20–30 m s$^{-1}$ range. The ARMAR measurements at nadir do not show such a saturation; $\sigma^0$ is a weakly decreasing function of wind speed even at high winds. The refer-
ence $\sigma^o$ found by the new method was compared with $\sigma^c$ found from the standard technique, allowing the effect of the constant surface wind assumption in the standard SRT to be assessed. The mean difference between the Doppler and standard methods is small ($\leq 1$ dB), while the rms difference is 0.9–2.6 dB. The assumed PIA standard deviation of at least 1 dB used in the TRMM PR rain retrieval algorithm is reasonably consistent with the rms differences found here.

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