Microwave Emission Brightness Temperature Histograms (METH) Rain Rates for Climate Studies: Remote Sensing Systems SSM/I Version-6 Results

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ABSTRACT

A satellite microwave emission brightness temperature histograms (METH) technique has been applied to Special Sensor Microwave Imager (SSM/I) data taken on board the Defense Meteorological Satellite Program (DMSP) satellites and preprocessed by Remote Sensing Systems (RSS) Co. to produce 21 yr (July 1987–present) of oceanic rainfall products. These rain products are used as input to the Global Precipitation Climatology Project (GPCP) rain maps. Analysis of the METH product using SSM/I version-4 (V4) data shows jumps in vertically polarized 19-GHz brightness temperatures that are attributed to changes in DMSP satellites. A version-6 (V6) SSM/I that corrects for intersatellite differences was released by RSS in 2006. The jumps in the time series are reduced, with most of the changes occurring in the early part of the DMSP F13 data. The bias between RSS V6 and V4 of brightness temperature at 19 and 22 GHz is less than 0.5 K. METH rain rates were reprocessed using V6 data and were analyzed. The 20-yr global mean difference between the METH V4 and V6 is less than 0.3%, with differences as large as 3% in individual years. Trend analyses show increases in the oceanic rain belts, such as the intertropical convergence zone and the South Pacific convergence zone, and in the Bay of Bengal. These rain-rate trends, from both linear trend analysis and empirical mode decomposition analysis, are comparable to the version-2 GPCP analyses but are smaller than those found in the unified microwave ocean retrieval algorithm.

1. Introduction

Oceanic precipitation is a major component of the hydrological cycle and climate processes. The oceans cover about 70% of the earth’s surface where most of the freshwater exchange occurs. The major driving force for atmospheric circulations also comes from the latent heat release associated with precipitation processes. Global precipitation is linked to the surface energy budget through evaporation, which occurs mostly over oceans. The understanding of trends and variability of oceanic precipitation and other rainfall-derived parameters is important for climate change and variability. To examine the response of the hydrological cycle to climate change, analyses of long-term hydrological-variable data are needed. Because of the scarcity of gauge networks, remote sensing data are the only means available for monitoring precipitation over the global oceans. The impact on the hydrological cycle associated with climate changes and global warming has been investigated in the past decades. There are general agreements on the trend of global temperature, but there is less consensus on changes in global precipitation (Folland et al. 2001; Hegerl et al. 2007; Karl and Trenberth 2003; Allen and Ingram 2002; Gu et al. 2007). Wentz et al. (2007) showed a 1.4 ± 0.5% increase in global precipitation and a 7% increase in the total amount of water in the atmosphere in response to a 1°C change in surface temperature from satellite observations. Their oceanic precipitation data are from the unified microwave ocean retrieval algorithm (UMORA) applied to Special Sensor Microwave Imager (SSM/I) data on board the Defense Meteorological Satellite Program (DMSP) satellites.
(Hilburn and Wentz 2008). Based on global rain maps produced by the Global Precipitation Climatology Project (GPCP), Adler et al. (2008) showed a 2.3% increase in global precipitation per 1°C increase in surface temperature, although their results are sensitive to the region and period of analysis. Associated with the warming, general circulation model results also point to an intensification of the hydrological cycle, which is supported by observations of the intensification of the Hadley circulation (Cess and Udelhofen 2003; Chiu and Xing 2004; Chiu et al. 2008). As part of the GPCP, the Polar Satellite Precipitation Data Center (PSPDC) provides monthly oceanic rainfall to the merging of the GPCP global product (Huffman et al. 1997). Note that the current version of PSPDC ocean rain maps is based on the so-called version-4 (V4) SSM/I data provided by Remote Sensing Systems (RSS) Co. In 2006, RSS released Version 6 (V6) of the SSM/I data and suggested its use for climate analysis. Because RSS is ending production of the V4 data, it is important to document the differences between the V4 and V6 results for climate trend and variability studies. The technique and data are described in section 2. The differences between V4 and V6 and results from linear trend analyses are discussed in section 3. Section 4 describes empirical mode decomposition (EMD) analyses applied to both datasets to quantify the trend and climate variability. The results are compared with UMORA and GPCP version 2 (V2) to show the effect of algorithm differences on trends. Section 5 summarizes our results.

### 2. Data

The dataset used in this study is the oceanic monthly rain rates produced at the GPCP PSPDC using the microwave emission brightness temperature histograms (METH) technique (Wilheit et al. 1991; Chang and Chiu 1999; Shin and Chiu 2008). The 20-yr (January 1988–December 2007) oceanic data are derived from the SSM/I on board the DMSP satellites. The DMSP satellites are in sun-synchronous orbits at altitudes of about 850 km. The orbital characteristics of DMSP satellites are shown in Table 1. Two monthly data products are available: a 5° × 5° monthly dataset for the ascending and descending DMSP satellite passes separately, covering the area from 50°N to 50°S, and a 2.5° × 2.5° monthly dataset covering 65°N–65°S. The data can be obtained from the GPCP PSPDC Internet site (http://gpcp-pspdc.gmu.edu). Although all DMSP satellite data are processed and are available from GPCP PSPDC, the version-2 GPCP monthly product is based on F8 for July 1987–December 1991, F11 for January 1992–April 1995, and F13 from May 1995 to the present to maintain homogeneity over the SSM/I product of the GPCP data record (Adler et al. 2003).

The METH technique for estimating monthly oceanic rain rates is based on a relationship between brightness temperature and rain rate ($T_b$–$R$) that is derived from radiative transfer calculation using a cloud model (Wilheit et al. 1991). In the cloud model, a Marshall–Palmer distribution (Marshall and Palmer 1948) of raindrops as a function of rain rate is assumed from the ocean surface to the freezing level (FL; 0°C) in the atmosphere. In addition, a constant lapse rate of 6.5°C km$^{-1}$ and a relative humidity that increases linearly with height from 80% at the ocean surface to 100% at the FL and above are assumed. This technique uses a linear combination of 2 times the 19-GHz vertical polarization minus 22-GHz vertical polarization to mitigate the effect of water vapor by minimizing the dependence of the brightness temperature on water vapor at low rain rates. The histograms of the combination channel $T_b$ over each space and time box are computed. The computed $T_b$ histograms are fitted iteratively to a rain-rate distribution (a mixed lognormal distribution) by using the $T_b$–$R$ relationship. The fitting of the $T_b$ histogram is based on three moments of the distributions (mean, variance, and skewness). At each iteration, all parameters—the mean $T_b$ and standard deviation $\sigma_0$ of the $T_b$ for the nonraining portion of the $T_b$ histogram, the rain fraction $p$, and the mean and standard variation of the conditional rain rates—are modified to minimize the errors between the estimated and assumed moments. Mean $T_0$ is also used to adjust the difference between the observed and computed $T_b$ histograms. The FL is considered as a proxy of

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1 METH V6 oceanic rainfall has been used in GPCP V2.1 starting from September 2008 (Huffman et al. 2009).

### Table 1. The orbital characteristics and the temporal coverage of the DMSP satellites (http://nsidc.org/data/docs/daac/f13_platform.gd.html). Orbit inclination is 98.8° for all three satellites.

<table>
<thead>
<tr>
<th>Satellite</th>
<th>Ascending time (at launch)</th>
<th>Alt (km)</th>
<th>Eccentricity</th>
<th>Orbital period (min)</th>
<th>Temporal coverage</th>
</tr>
</thead>
<tbody>
<tr>
<td>F8</td>
<td>0615</td>
<td>851–832</td>
<td>0.001 32</td>
<td>101.8</td>
<td>Jul 1987–Dec 1991</td>
</tr>
<tr>
<td>F11</td>
<td>1811</td>
<td>878–841</td>
<td>0.001 29</td>
<td>101.9</td>
<td>Jan 1992–Apr 1995</td>
</tr>
<tr>
<td>F13</td>
<td>1742</td>
<td>856–844</td>
<td>0.000 83</td>
<td>102.0</td>
<td>May 1995–present</td>
</tr>
</tbody>
</table>
the columnar humidity content. The altitude of the FL is
determined using the mean of the upper 99th percentile
of the vertically polarized 19- and 22-GHz brightness
temperatures. Biases due to nonuniform beam coverage
are corrected using the physical–empirically derived
beamfilling correction factor (Wang 1997; Chiu et al.
1990).

From the \( T_b - R \) relationship, \( T_b \) increases with rain rate
at the lower rain rates (emission regime), levels off, and
decreases for higher rain rates because of scattering of the
hydrometeors. Sensor biases between satellites can con-
tribute to biases in rainfall estimates. Since \( T_b \) increases
with rain rate for the frequently observed rain-rate range,
a low bias in \( T_b \) would result in a lower rain rate, for
a fixed \( T_b - R \) relationship. However, the temperature
biases are compensated by changes in \( T_0 \); hence the
technique is self-calibrating, to a first approximation.

3. Differences between V4 and V6 data

There are differences in the input \( T_b \) and the algorithm
versions between RSS V4 and V6 as discussed below. For
\( T_b \), “In V4, F8 data was accepted as is. . . . the bias of
subsequent satellites in relation to F8 [was calculated]. In
V6, all satellites were averaged to use as the base, and then
the bias was calculated for each satellite in relation to that
average. . . . A \( T_b \) difference in the range of 0.5° between
V4 and V6 can be expected.” (F. Wentz via S. Tremble,
RSS, 2008, personal communication). The versions of the
METH SSM/I products correspond to that of the RSS \( T_b \)
products. No METH products using V5 RSS SSM/I were
produced. There are calibration problems with the RSS
V5, which shows a spurious trend in surface wind. The \( T_b 
\) data were recalibrated in V6 to make the \( T_b \) consistent
among the satellite platforms that include SSM/Is (Re-
 mote Sensing Systems 2009). The spurious wind trend was
removed in RSS V6, and the oceanic products are now
called UMORA (Hilburn and Wentz 2008). Additional
features of the V6 \( T_b \) data include a land/sea flag and a sea
ice flag attached to each pixel.

There are three major differences in the METH V4 and
V6 algorithm. First, for quality controls of \( T_b \) data, all
pixel \( T_b \) values of less than 50 K or more than 350 K for
all channels are first identified. In the V4 algorithm, the
whole scan is rejected if any one pixel from any channel is
found to be outside this range. In V6 only the out-of-range
pixels are rejected. Second, a land/sea mask at 0.25° res-
olution generated from S’ gridded elevations/bathymetry
for the world (ETOPO5: NOAA 1988; see online at http://
www.ngdc.noaa.gov/mgg/global/etopo5.html) is used
in V4. Pixels that are within one pixel width from land
are not included. In the V6 \( T_b \) data, each pixel is tagged
as either land or sea, and only the sea pixels are included.

The third aspect involves the screening out of sea ice
areas. In V4, the algorithm uses the monthly climato-
logical values of sea ice to locate and screen out sea ice
regions. The V6 algorithm uses the RSS V6 sea ice flag
to screen out pixels identified as sea ice.

To examine the differences between V4 and V6 SSM/I
\( T_b \), Figs. 1a and 1b show the time series of the average \( T_b 
\) at vertically polarized 19 GHz (Tb19v) and 22 GHz
(Tb22v) for V4 and V6, respectively. The \( T_b \) averages
were calculated from \( T_b \) histograms of a 2.5° × 2.5° grid
product for oceanic areas from 50°N to 50°S. The do-
main \( T_b \) averages are 201.57 and 201.81 K for Tb19v,
and 229.01 and 229.11 K for Tb22v, respectively, for V4
and V6. The 20-yr average RSS V6 is higher than that of

\[<\text{fig1.jpg}>\]
V4 by 0.24 K at Tb19v and 0.1 K at Tb22v. Figure 1c shows the time series of the difference (V6 – V4) for Tb19v and Tb22v. Temporal variations of the difference (V6 – V4) are larger for Tb19v (0.2 K) than for Tb22v (0.04 K). For both the DMSP F8 and F11 satellites, V6 is higher than V4 for both channels. For the F13 satellite, the V6 Tb19v is lower than V4 during the first 2 yr of the F13 satellite era and is higher afterward. For Tb22v, V6 is generally higher than V4 during the northern winter and is lower during the northern summer for F8 and F11. For F13, the differences show two peaks, around February and October.

The Tb differences between the versions are 0.41, 0.46, and 0.07 K for Tb19v and 0.14, 0.24, and −0.04 K for Tb22v for F8, F11, and F13, respectively. The differences in the combination channel for F8 are less than those for F11. The differences for F13 are the smallest, probably because of its long data record used in the calibration of long-term average. As a result, the jumps in Tb19v between satellites are smaller in V6 than in V4 (Fig. 1a).

The time series of oceanic rain rates, the mean of combination channel brightness temperature for the nonraining portion T0, and the FL are shown in Fig. 2. The version change affects the Tb and hence T0. Figure 3 shows the differences of domain-average oceanic rainfall, T0, and FL between V6 and V4. The mean rain rates computed from METH V4 and V6 data are 2.96 and 2.97 mm day$^{-1}$, respectively, which is a <0.3% difference. In general, the absolute rainfall difference between the two versions is less than 0.05 mm day$^{-1}$ (≈1.5%), except for the first 2 yr of the F13 satellite era in which V6 rain rates are smaller than V4 by about 0.09 mm day$^{-1}$ (≈3%).

The T0 averages for the 20 yr are 172.00 and 172.37 K for V4 and V6, respectively. Chokngamwong and Chiu (2006) showed jumps in time series of T0 in the V4 data that were attributed to satellite changes—in particular, the transition from F11 to F13. Even though the jumps still exist in V6, they are smaller than that of V4. Overall, the decreasing trend in T0 becomes more apparent. Linear trend analyses of T0 show decreasing trends of 0.36 and 0.60 K (10 yr)$^{-1}$ for V4 and V6, respectively. The changes in T0 basically follow the changes in Tb19v because of the fact that T0 is weighted by 2 times Tb19v. The relatively large changes in Tb19v in the first 2 yr of F13 data result in lower oceanic rainfall, lower T0, and a higher FL for V6 during this time period.

The domain-average FL decreases by approximately 0.2% (4.30 and 4.29 km for V4 and V6, respectively). The METH technique estimates the emission from the rain volume; hence, an underestimate of the FL would result in an overestimate of the rain rate for fixed T0, and vice versa (Wilheit et al. 1991; Chiu and Chang 2000).

The lower FL would suggest a higher rain rate for V6. Biases in both Tb19v and Tb22v introduce a bias in T0. The increase in T0 in the T0–R relation is less than the increase in T0 except for zero rain rate (see the appendix). If the T0 increase in V6 is the same for all rain rates, a higher rain rate would be retrieved. Both the FL and T0 factors are at work in shaping the difference between V6 and V4 rain rates. The rain rate is more sensitive to FL changes than to T0 changes, as evidenced in the higher V6 rain rate and lower FL overall. However, during the first 2 yr of F13, increases in FL and decreases in T0 both result in lower V6 rain rates.

Figure 4 shows the distribution of seasonal rain-rate biases. Although the biases are small and the grids are
scattered, a general pattern can be summarized as follows: In general, there are positively biased (V6 > V4) grids in the intertropical convergence zone (ITCZ) area and negatively biased grids in the “Maritime Continent,” most of which are coastal grids. These negative biases may be related to the use of the land/sea mask. In V4 the land mask is based on a 0.25° grid created from ETOPO5. In V6, the pixels are classified using the RSS-supplied land/sea mask. The result is fewer samples for V6 than for V4 for these grids. Some of the pixels included in V4 may be land contaminated. During the northern winter (December–February), positively biased grids appear in the northern high latitudes in the sea ice routes. The positive bias may be associated with the different application of a sea ice mask. The monthly climatological sea ice mask is used in V4 whereas the pixels are classified as sea ice according to the RSS-supplied classification in V6. The appearance of a larger number of grids with positive bias (V6 > V4) in the North Pacific and North Atlantic Oceans may explain the seasonal cycle of the domain-average rain rate in Fig. 3a.

In summary, the difference between V6 and V4 is noted in a lower FL of 0.01 km (0.2%), a higher $T_0$ of 0.37 K (0.2%), and a higher rain rate of 0.01 mm day$^{-1}$ (0.3%). Although there are scattered grids that show biases, the V4 and V6 spatial patterns are in general similar. The general trend patterns of the V4 and V6 rain rates are also similar and are discussed next.

4. Linear trend and empirical mode decomposition

To quantify the trends, Fig. 5 shows the spatial pattern of the trend from linear regression analysis for both V4 and V6. The slope of the linear regression, in units of millimeters per day per month, shows the trend. The linear trend pattern for both V4 and V6 is very similar. Significant increasing trends mostly occur in the ITCZ, South Pacific convergence zone, and Bay of Bengal. However, small regions of decrease are found near northeastern coastal Australia. Analyses of zonal mean rain rates show a significant increasing trend of about 4% over the 0°–10°N ocean, with rates of 0.234 and 0.238 mm day$^{-1}$ (10 yr)$^{-1}$ for V4 and V6, respectively. The linear trends for the whole domain (50°N–50°S), the tropics (25°N–25°S), and the region of most noted increase (0°–10°N) are summarized in Table 2. Both the mean and the linear trend for these regions are similar between V4 and V6; the differences in the means are less than 0.3%. The magnitudes and the percentage change (over the 20-yr period) increase from the domain average to the tropics. The highest increasing trend of about 4% for the 0°–10°N region is significant at the 95% level by testing the significance of the slope of the linear regression.

The trends are dependent on the length of the time series and hence may be nonlinear or nonstationary. To illustrate the “trends” more clearly, an EMD analysis (Huang et al. 1998) is applied to the time series of the domain-average rain rates and rain rates at 0°–10°N. The EMD is a useful tool for analyzing nonlinear and nonstationary processes on the basis of the local characteristic time scales of the data. It decomposes the time series into sequences of intrinsic mode functions (IMF) of increasing time scales. The existence of a trend is indicated if the last IMF (with the longest time scale) is monotonically increasing or decreasing. Figure 6 shows

(a)

(b)

(c)
the last IMFs of the domain-average rain rates of 50°N–50°S and 0°–10°N for V4 and V6, respectively.

Over the domain (50°N–50°S), the last IMF of V6 shows a monotonic increase of about 0.018 mm day$^{-1}$ over the 20-yr period (computed from the difference in the end points of IMF 7), or 0.009 mm day$^{-1}$ (10 yr)$^{-1}$. The trend in V4 is somewhat larger: 0.052 mm day$^{-1}$ (10 yr)$^{-1}$. For 0°–10°N, V4 shows an increase of 0.153 mm day$^{-1}$

Fig. 4. Seasonal bias (V6 − V4) of oceanic rain rates for the 20-yr period. Grids with less than 50% of the samples (30 months) are excluded. Grids with biases of <0.25 mm day$^{-1}$ and missing grids are colored gray. Here, DJF indicates December–February, MAM is March–May, JJA is June–August, and SON is September–November.
and V6 shows an increase of 0.110 mm day\(^{-1}\) (10 yr\(^{-1}\)). The trend results obtained from EMD analyses are smaller than those obtained from linear trend analyses, indicating that other climatic variability may have contributed to the linear trend.

Although the trends from the nonlinear EMD analyses are smaller than the linear trends, they all show increasing trends—in particular, in the region 0\(^\circ\)–10\(^\circ\)N. This decadal increase is consistent with the general circulation model results and satellite data analyses that show a net increase in rainfall over tropical oceans and intensification of the atmospheric Hadley circulation.

To show the algorithm effect on rain rates, the mean rain rates, linear trends, and percentage changes from means for the same regions are computed for UMORA and GPCP V2. The results are included in Table 2. The means, trends, and percent change are also similar between V6 and V4 and are slightly lower for GPCP V2 for all of these regions. The trends are about 1.6% and 3.1% for the globe (50\(^\circ\)N–50\(^\circ\)S), 1.8% and 4.0% for the tropics, and 4.4% and 7.0% for 0\(^\circ\)–10\(^\circ\)N for METH V6 and UMORA, respectively, indicating higher trends in all regions as determined from UMORA.

### 5. Summary and discussion

The brightness temperature data taken by the SSM/I on board the DMSP F8, F11, and F13 satellites for 1988–2007 are examined to quantify the differences between the RSS V6 and V4 \(T_b\) data. Over the period, V6 \(T_b\) are slightly higher than those of V4 by 0.25 and 0.1 K, respectively, for \(T_b1\) and \(T_b2\). However, there are

| Table 2. Comparison of mean rain rates and linear trends among V4 METH, V6 METH, V2 GPCP, and UMORA data for different spatial domains over the period 1988–2007. Units are millimeters per day for the mean, millimeters per day per 10 yr for the trend, and total percent per 10 yr for percentage change. For comparison, trend for GPCP for 1979–2006 (Adler et al. 2008) for 25\(^\circ\)N–25\(^\circ\)S is 0.059 mm day\(^{-1}\) (10 yr\(^{-1}\)). |
|---|---|---|---|---|---|---|
| | 50\(^\circ\)N–50\(^\circ\)S | 25\(^\circ\)N–25\(^\circ\)S | 0\(^\circ\)–10\(^\circ\)N | 1988–2007 | Mean | Trend | Percent change | Mean | Trend | Percent change | Mean | Trend | Percent change |
| V4 METH | 2.96 | 0.039 | 1.32 | 3.15 | 0.053 | 1.68 | 5.39 | 0.234 | 4.34 |
| V6 METH | 2.97 | 0.046 | 1.55 | 3.15 | 0.057 | 1.81 | 5.42 | 0.238 | 4.39 |
| V2 GPCP | 2.96 | 0.023 | 0.78 | 3.03 | 0.050 | 1.65 | 5.11 | 0.216 | 4.23 |
| UMORA | 2.64 | 0.083 | 3.14 | 2.97 | 0.119 | 4.01 | 5.40 | 0.376 | 6.96 |

[FIG. 5. Linear trend (mm day\(^{-1}\) month\(^{-1}\)) of V4 and V6 oceanic rain rates during the 20-yr period. As an example, a trend of 0.008 mm day\(^{-1}\) month\(^{-1}\) corresponds to 0.96 mm day\(^{-1}\) (10 yr\(^{-1}\)).]
intersatellite differences, with the Tb19v showing much larger variability in the difference (V6 − V4). For Tb19v, the jumps in the V4 time series between F11 and F13 satellites are smaller in V6. Large version differences are found in the early part of the F13 data.

Differences in rain rate can be attributed to both the FL and T0, the brightness temperature for no rain. Examination of the seasonal differences shows scattered positive-biased (V6 > V4) grids in the ITCZ region and in the sea ice region. Negative-bias grids are found in the Maritime Continent, mostly in coastal regions. They are attributed to the difference in the application of land/sea mask and sea ice mask in the different versions. In METH V4, a land mask at 0.25° resolution was created using ETOPO5 data and the surface types (ocean or land or mixed) are specified at a 0.5° resolution. In V6, the land surface types are specified for every pixel and hence provide an improvement in land contamination of Tb for pixels close to land.

Analysis of GPCP global rain maps shows an increasing rainfall trend in the tropical ocean, with opposite trend over land (Gu et al. 2007). The spatial patterns of the linear trends of oceanic precipitation are very similar. Table 2 summarizes our results of trends from linear regression. Trends of 0.039 (1.32%) and 0.046 (1.55%) mm day$^{-1}$ (10 yr)$^{-1}$ are found over 50°N–50°S for V4 and V6, respectively, whereas the trend from EMD analysis is smaller. Adler et al. (2008) found an increase of 0.0588 mm day$^{-1}$ (10 yr)$^{-1}$ over the tropical oceans of 25°N–25°S. This is comparable to both METH V4 and V6 results. We also found a significant increase of about 4% from both linear regression analysis and EMD analysis for the zonal belt 0°–10°N. This increasing trend in tropical precipitation in the ITCZ is consistent with the results of Kumar et al. (2004) from atmospheric general circulation model simulations, which show a net increase in rainfall over tropical oceans, and the intensification of the atmospheric Hadley circulation from analysis of satellite data (Cess and Udelhofen 2003; Chiu and Xing 2004).

A comparison was made among METH V6 and V4, GPCP V2, and UMORA. Our results show that the means and trends in METH V6 and V4 and GPCP V2 are very similar. UMORA shows smaller means in the global domain and tropics, but trends are larger than that of V6 by almost a factor of 2.

Based on our analyses, we conclude that means and trends in METH V6 are generally similar to V4. Improvements in V6 result from the use of an improved land/sea mask, a sea ice mask, and the nonexclusion of good T5 data on the same scan that adds to the samples. Because of the discontinued production of the RSS V4 T5 data, we therefore recommend switching from V4 to V6 for the processing of the GPCP PSPDC METH rain rates, as has been recommended by RSS for processing of climate-scale products.

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**APPENDIX**

**Changes in Retrieved Rain Rate due to T5 Biases between V4 and V6**

We consider changes in the T5–R relation due to changes in Tb19v and Tb22v. According to Figure A1 in Wilheit et al. (1991), increases in Tb19v decrease the FL while increases in Tb22v increase the FL. When the increase in Tb19v is larger than that in Tb22v, lower FL can be expected. The larger increase is due to two
factors: the bias in Tb19v is larger than that in Tb22v and the combination channel weights more (2 times) on Tb19v than on Tb22v. Note that the scales in their Fig. A1 are the same for both axes. Because the FL is coupled directly to the rain rate, the FL effect is expected to be the dominant effect.

As discussed in section 2, the use of the parameter $T_0$ absorbs a number of approximations and hence the algorithm is self-calibrating, to a first approximation. The relation between the combination-channel brightness temperature and rain rate is approximated as [Wilheit et al. (1991), their Eqs. (4) and (5)]

$$T(r) = T_0 + (285 \text{ K} - T_0)(1 - e^{-\frac{r}{r_c}}) - 3.5r^{1/2}$$

and

$$r_c = \frac{25}{F^{1/2}},$$

(A1)

where $T$ is the combination-channel brightness temperature, $r$ is rain rate, $T_0$ is the $T$ at zero rain rate, and $F$ is the FL in kilometers for the $T_{0-R}$ relation. Consider a small change in $T_0$. The corresponding change in $T$ in the $T_{0-R}$ relation is

$$\delta T = \delta T_0[1 - (1 - e^{-r/c})] = \delta T_0 e^{-r/c}.$$ 

For an increase (decrease) in $T_0$, the corresponding increase (decrease) in $T$ is the same at zero rain rate, but the increase (decrease) is less than $T_0 \left(e^{-r/c} < 1\right)$ as the rain rates increase. Hence for range-independent biases in Tb19v and Tb22v, the observed combination channel $T_b$ will in general be lower than the $T$ in the $T_{0-R}$ relation; hence, a lower retrieved rain rate can be expected. The effect is expected to be secondary to the FL effect as discussed earlier.

REFERENCES


