TRMM Precipitation Radar and Microwave Imager Observations of Convective and Stratiform Rain over Land and their Theoretical Implications

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Abstract

Observations of brightness temperature, T_b , made over land regions by the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) radiometer are analyzed with the help of nearly simultaneous measurements of the vertical profiles of reflectivity factor, Z , made by the Precipitation Radar (PR) onboard the TRMM satellite. Furthermore, this analysis is done separately over convective and stratiform rain regions. This examination reveals a clear relationship between TMI and PR data. Possible explanation for this relationship is explored with the help of radiative transfer calculations. With this approach, we demonstrate that the 85 GHz observations of TMI can be simulated crudely from the observations of Z . However, the 37 and 19 GHz observations are not as well simulated, possibly because of horizontal non-uniformity in the hydrometeor distribution in the broad footprints of these channels and contamination introduced by land-surface emissivity. On the other hand, from TMI and PR observations, we find that the brightness temperature difference ($T_{19}-T_{37}$) minimizes these sources of error. Our simulations of ($T_{19}-T_{37}$) over convective rain regions are in reasonable agreement with this finding. This investigation indicates that the TMI 85 GHz channel yields the best information about rain over tropical land, because it has minimal surface contamination, strong extinction, and a fine footprint. The brightness temperature difference ($T_{19}-T_{37}$) can supplement the information given by the 85 GHz channel.

1. Introduction

Observational studies of convective and stratiform rain reveal that these two types

of rain are associated with distinctly different types of hydrometeors in a vertical column of the atmosphere (e.g., Houze 1993, 1997). This is mainly because of differences in the dynamics and the physical processes associated with these two rain types. This observational background helps us to investigate the satellite measurements from the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager

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(TMI) radiometer and Precipitation Radar (PR) over convective and stratiform rain regions.

Onboard the TRMM satellite, the TMI and PR produce nearly simultaneous observations over ocean and land under different weather conditions. The brightness temperatures, T_b s, measured by TMI represent upwelling microwave radiation that has undergone interaction with atmospheric oxygen, water vapor, cloud liquid water, and liquid and frozen hydrometeors in a vertical column of the atmosphere. On the other hand, the measurements of reflectivity factor Z (from hereafter referred to as reflectivity) made by PR near 13.8 GHz depend upon the backscatter of microwave radiation, and have a vertical resolution of 250 m. Reflectivity observations near the surface can yield an estimate of the rain rate at that level when a suitable particle size distribution (PSD) is assumed. Based on the above discussion, one expects to find a relationship between the vertical profiles of Z observed from PR and the T_b measurements of TMI. The objective of this study is to reveal such a relationship with the help of TMI and PR data, and then explain that relationship theoretically.

Before the launch of the TRMM satellite, a feasibility study made by Haddad et al. (1997) showed that the vertical profile of rain could be deduced from a combination of the PR and TMI data. In another study, in order to understand the link between microwave radiometer and radar observations, Bennartz and Petty (2001) performed an analysis combining the brightness temperatures observed by Special Sensor Microwave Imager radiometer (flown on Defense Meteorological Satellite Program satellites) and measurements of reflectivity made by a surface radar over the Baltic Sea near 57°N. Their study reveals the effect of variable size distribution and density of precipitating ice particles on microwave T_b s.

Viltard et al. (2000) examined the importance of the PSD in linking the PR data with the TMI measured T_b s over oceans utilizing radiative transfer simulations. The vertical profiles of the water content in the atmosphere, needed in those simulations, were selected from a table of profiles generated with the help of a cloud resolving model. However, the selected profiles of the water content were not in complete

agreement with the PR reflectivity profiles. Adopting that approach, Viltard et al. were able to show crudely the significance of PSD over ocean.

In this study, we probe into the connection between the TRMM PR and TMI observations over land following a different approach that emphasizes the measurements of these two instruments. First, we have compared the vertical profiles of Z given by the PR with the T_b s in the 19, 37, and 85 GHz—i.e., T19, T37, and T85—measured by the TMI. This is done separately for convective and stratiform rain areas, which are discriminated by the PR as indicated in the 2A25 TRMM product (see Iguchi et al. 2000). This analysis of observations is useful to reveal the inherent relationship between the PR and TMI data. Then, hydrometeor profiles that are consistent with the Z profiles are formed with guidance from several past studies of the microphysical structure in convective and stratiform rain regions. Utilizing these hydrometeor profiles, radiative transfer calculations are made to simulate the T_b s observed by TMI. From this investigation, we can infer the relative importance of different channels of the TMI radiometer for the purpose of rain retrieval over land.

2. Correspondence between TMI and PR observations of convective and stratiform rain over land

The near-surface PR observations have a horizontal resolution of about $4.3 \text{ km} \times 4.3 \text{ km}$. These PR data are arranged uniformly at a spacing of about 4.3 km both along and across scan lines. On the other hand, the footprints of the TMI 85 GHz vertical and horizontal polarization channels are approximately $5.0 \text{ km} \times 7.0 \text{ km}$, and are separated by about 5.0 km along scan lines and about 14 km across scan lines. This scan pattern degrades the effective resolution of the TMI 85 GHz channel compared to that of the PR. The TMI radiometer contains additional channels in vertical and horizontal polarization near 10, 19, and 37 GHz that have a resolution of about 40, 20, and 10 km, respectively. Furthermore, TMI has a 21 GHz channel in vertical polarization that has a resolution similar to that of 19 GHz. In this study, we are interested in scrutinizing the information contained in the vertical profile of

Fig. 1. Maps of **a)** T19, **b)** T37, **c)** T85, **d)** (T19-T37), and **e)** PR near-surface rain rate for a mesoscale convective system observed by the TRMM satellite on 5 June 1998 over the Southeast United States. Note that the brightness temperatures T19, T37, and T85 represent the vertical polarization measurements.

Z measured by the PR and the Tbs measured by TMI in the 19, 37, and 85 GHz channels. The TMI 10 and 21 GHz channels are not included in this investigation. This is because the 10 GHz channel has a relatively large footprint, and the information given by the 21 GHz channel is very similar to that of 19 GHz. Also, we consider only the vertical polarization measurements made by TMI, because they tend to have less surface contamination compared to horizontal polarization measurements.

In Figs. 1a–e, we show maps of T19, T37,

T85, and (T19-T37), as well as a map of PR near-surface rain rate, for a mesoscale convective system (MCS) that occurred over the Southeast United States on 5 June 1998. From these figures, we note that the spatial distribution of PR rain rate compares well with that of T85, but poorly with that of T19 or T37. This is mainly because of the strong extinction due to hydrometeors in the 85 GHz channel compared to that in the long wavelength channels near 37 and 19 GHz. The surface emission present in these long wavelength channels can degrade

the rain information. However, a map of the difference, (T19-T37), which reduces this surface contamination, produces a good comparison with the PR rain rate patterns. Such information given by the microwave radiometer and the radar is noticed over widely different regions of the tropics.

We show in Fig. 2a a vertical cross section of the radar reflectivity taken from TRMM PR along a sub-satellite track. This cross section, which starts at (35.1°N, 90.6°W) and ends at (35.0°N, 87.6°W), corresponds to the same MCS depicted in Figs. 1a–e. The PR reflectivity below the 1 km level is not shown in this figure to avoid problems related to signal clutter arising from surface topography. We note from Fig. 2a that the leading edge of the storm, with its tall convective towers, is near 88°W. High reflectivity cores (>30 dBZ) that extend well above the freezing level (~5.0 km) characterize these convective towers. To the west of these convective towers, from 88.6°W to 90.3°W, and above the freezing level, we notice an extensive anvil cloud with an almost horizontally uniform pattern of reflectivity. Below the freezing level, underneath the anvil cloud, there is a bright band that is about 0.5 km in thickness. Strong reflectivity in the bright band is attributed to large, low-density melting snow flakes, while the strong reflectivity above the freezing level in the cores of the convective towers is attributed to small, water-coated dense graupel (Bauer et al. 2000). Near the surface, there is intense convective rain in the convective towers, while there is relatively light stratiform rain underneath the anvil (see Fig. 2c). Notice there is a small region referred to as the Bounded Weak Echo Region (BWER) between 88.4°W and 88.5°W in the layer from 2 km to 6 km, which is at the origin of the anvil cloud. Such a BWER has been described in earlier studies with the help of observations made by ground-based radars (see Houze 1993). This description of the reflectivity pattern and rain rate associated with a MCS is helpful in relating the TRMM PR observations with those of TMI.

In Fig. 2b, the brightness temperatures measured by TMI in the 85, 37 and 19 GHz channels are shown along the sub-satellite track to correspond to the cross section of PR reflectivity shown in Fig. 2a. In the response of the 85 GHz

channel, we find there is a strong scattering depression due to relatively dense ice particles (Wu and Weinman 1984) where the convective towers are present, but some of the individual towers are not well resolved. The anvil in the stratiform region produces weaker scattering depression. Over the entire cross-section domain, the response of the 37 GHz channel is similar to that of the 85 GHz channel, except that the scattering depression in 37 GHz is not as strong. The response in the 19 GHz channel is the weakest of the three channels. The inter-relationship between TMI measured T85, T37, and T19, and the PR measured Z and R, is only crudely reflected in Figs. 2a–c, because of differences in the resolution of the TMI and PR data.

To compare the TMI and PR data sets, we have averaged them to the TMI 19 GHz field of view (*fov*), which has a size that is comparable to that of Cbs. As indicated earlier, the TMI scan data are not uniformly spaced. For this reason, to reduce the effect of non-uniformity in TMI data, we have applied a simple spatial averaging procedure.

The TMI has a conical scanning geometry, while the PR has a cross-track scanning geometry. These differences in scanning geometry can introduce spatial mismatch between PR and TMI data (Hong et al. 2000). The PR data near the surface is generally used to deduce rain information. On the contrary, the 85 GHz temperature responds to the hydrometeors in the whole column of the atmosphere, with special emphasis on ice hydrometeors above the freezing level (Wu and Weinman 1984). Thus, differential advection of hydrometeors between the upper and lower levels can create spatial mismatch between the 85 GHz data of TMI and PR near-surface rain observations. Also, the hydrometeors originating at higher altitudes take several minutes to reach the surface. For these reasons, averaging the data of each instrument to the 19 GHz *fov* will reduce the problems involved in comparing the information content of these two datasets. Similar arguments also apply to the 37 GHz data. However, compared to the 85 GHz, the 37 GHz channel responds to lower layers of the atmosphere, while the 19 GHz channel is sensitive to layers even further below. This is due to reduced ice scattering in these longer wavelength

Fig. 2. Vertical cross-sections of **a)** PR reflectivity, **b)** TMI T19, T37, and T85, and **c)** PR near-surface rain rate for the MCS presented in Figure 1 (see text for details).

Fig. 3. Observed PR rain rate versus 85 GHz brightness temperature for **a)** *pixels* (see text) and **b)** averages of *pixels* over 20 K intervals of T85. The open circles denote observations from convective regions, while the filled circles represent observations from stratiform regions. Data for these plots are taken from TRMM observations of 20 MCS cases over land.

channels. In this study, we have assumed that simple averaging of the TMI 85 and 37 GHz data to the 19 GHz *fov* also minimizes this spatial mismatch. A similar averaging is also applied to the PR measured reflectivity, Z , at several discrete altitudes and to the near-surface rain rate, R . All of these TMI and PR data averaged to the 19 GHz *fov* are referred to from hereafter as *pixel* data.

The PR data show that rain rate can be highly variable within a 19 GHz *fov*. There can be about 30 PR footprints of $4.3 \text{ km} \times 4.3 \text{ km}$ in this *fov*. These 30 footprints can be sorted according to the PR classification as having stratiform rain or convective rain. In addition, a small number of these PR footprints could be rain free, or have rain that does not fit in the above two categories. Here we have adopted the PR classification of convective and stratiform rain given in the Version 5 of the 2A25 PR data sets (see Iguchi et al. 2000). Also, we assume PR-observed, near-surface rain rate is zero when it is less than 0.1 mmhr^{-1} . In this study, a given *pixel* is categorized as convective

when 50% or more of its area is covered by rain of convective type. This threshold of 50% is chosen because convective rain area tends to have a small scale. Similarly, a *pixel* is categorized as stratiform when 80% or more of its area is covered by stratiform type of rain. For the purpose of generating a set of TMI and PR data associated with such convective and stratiform *pixels*, we have utilized 20 MCS rain events observed by the TRMM satellite over diverse land regions of the tropics (Africa, Asia, Australia, North America, and South America). Note, to reduce contamination from wet land surfaces, the data where the polarization difference between the vertical and horizontal channels in the 19 GHz is greater than 5 K are eliminated.

In Fig. 3a, we present a plot of T85 and the corresponding PR rain rate that have been averaged to TMI 19 GHz *fov*. This data is gathered from the chosen 20 MCS events. From this plot, we see that the *pixels* characterized as convective by the PR have relatively high rain rates, and the *pixels* characterized as strati-

Table 1. Average of PR and TMI *pixel* observations (see text) within a given 20 K interval of T85 for **a)** convective and **b)** stratiform rain. The averages are calculated for the following parameters: T85; T37; T19; convective rain fraction (*C*); stratiform rain fraction (*S*); PR near-surface rain rate, *R*; and PR reflectivity, *Z*, at 1, 3, 4, 4.5, 6, 7, 9, and 11 km. The number of 20 km *pixel* observations in a given T85 interval is also shown in these tables.

a) Convective Rain Pixel Statistics																
	T85 Interval K	# of Pixs.	$\overline{T85}$ K	$\overline{T37}$ K	$\overline{T19}$ K	<i>C</i> %	<i>S</i> %	$\overline{R}$ mm hr	$\overline{Z}$ @ 1 km	$\overline{Z}$ @ 3 km	$\overline{Z}$ @ 4 km	$\overline{Z}$ @ 4.5 km	$\overline{Z}$ @ 6 km	$\overline{Z}$ @ 7 km	$\overline{Z}$ @ 9 km	$\overline{Z}$ @ 11 km
A	110-130	21	121.9	200.1	257.8	91	8	51.2	1.4E5	1.4E5	1.1E5	8.3E4	3.2E4	1.6E4	4.9E3	1.5E3
B	130-150	48	140.4	215.5	261.4	87	9	43.0	1.2E5	1.1E5	7.9E4	6.0E4	2.1E4	1.0E4	2.7E3	8.2E2
C	150-170	48	159.0	229.5	266.4	83	15	31.2	6.9E4	6.7E4	4.6E4	3.3E4	1.1E4	5.9E3	1.8E3	7.2E2
D	170-190	102	181.2	244.4	271.7	77	22	25.6	4.0E4	3.1E4	2.2E4	1.5E4	3.9E3	1.9E3	4.6E2	2.1E2
E	190-210	139	200.2	252.5	274.4	72	25	22.7	3.5E4	2.3E4	1.6E4	1.1E4	2.9E3	1.4E3	4.2E2	1.9E2
F	210-230	132	219.5	258.7	275.8	73	23	20.2	2.7E4	1.8E4	1.3E4	9.4E3	2.1E3	7.9E2	2.4E2	9.6E1
G	230-250	78	239.1	265.3	278.4	65	29	14.2	1.4E4	1.1E4	9.3E3	7.2E3	1.4E3	5.0E2	1.5E2	7.3E1

b) Stratiform Rain Pixel Statistics																
	T85 Interval K	# of Pixs.	$\overline{T85}$ K	$\overline{T37}$ K	$\overline{T19}$ K	<i>C</i> %	<i>S</i> %	$\overline{R}$ mm hr	$\overline{Z}$ @ 1 km	$\overline{Z}$ @ 3 km	$\overline{Z}$ @ 4 km	$\overline{Z}$ @ 4.5 km	$\overline{Z}$ @ 6 km	$\overline{Z}$ @ 7 km	$\overline{Z}$ @ 9 km	$\overline{Z}$ @ 11 km
A	190-210	201	204.4	257.9	275.9	6	94	6.0	4.3E3	4.7E3	4.9E3	1.5E3	2.5E2	1.7E2	8.3E1	3.2E1
B	210-230	587	220.4	260.4	277.2	4	96	5.9	3.9E3	4.4E3	6.2E3	3.3E3	2.7E2	1.6E2	6.8E1	1.9E1
C	230-250	668	240.4	265.1	279.3	2	97	4.2	2.2E3	2.5E3	4.2E3	2.6E3	1.8E2	1.1E2	3.7E1	6.6E0
D	250-270	562	257.2	271.2	280.7	1	95	2.0	7.8E2	9.4E2	1.6E3	9.1E2	1.0E2	6.5E1	1.3E1	1.9E0

form have generally low rain rates. A closer examination of the PR data shows that most of the convective *pixels* have rain rates greater than 10 mmhr^{-1} , even at warm brightness temperatures as high as 260 K. In the stratiform *pixels*, more than 97% have rain rates less than 10 mmhr^{-1} . There is a lot of scatter in the data, however, T85 decreases as *R* increases.

To enhance further statistical strength in the *pixel* data described above, we have performed additional averaging. First the PR and TMI *pixel* data are sorted according to the T85 value. Then in each 20 K interval of T85, averages of the TMI T19, T37, and T85 *pixel* data are calculated. Similar averages of the near surface rain rate given by PR and the reflectivity, *Z*, at individual altitudes are calculated. This has been done separately for convective and stratiform regions using the data from the chosen 20 MCS events. These averages of *pixel* data are presented in Tables 1a and 1b, and are represented with an overbar, i.e., $\overline{T19}$, $\overline{T37}$, $\overline{T85}$, $\overline{Z}$ and $\overline{R}$. Each row in these

tables represents a sequential 20 K interval in T85 and is identified alphabetically. The number of *pixels* that are averaged in each row is also shown in the tables. It may be pointed out that the standard deviation of all observations of *Z* at each altitude for a given T85 interval of 20 K has a magnitude that is comparable to the mean value itself. For each row in Tables 1a and 1b, the fractional area of convective rain (*C*) and stratiform rain (*S*) is indicated as a percentage. Also, since these two rain areas do not add up to 100%, we can infer a small area (<6%) with rain that does not fit into the *C* or *S* category, or that could be rain free.

Taking the *pixel* averages given in Tables 1a and 1b, we present in Fig. 3b the relationship between $\overline{R}$ and $\overline{T85}$ for the convective and stratiform regions. This figure shows clearly that when one averages the data in the above manner, the variability in the data is reduced, and a clear relationship between $\overline{R}$ and $\overline{T85}$ emerges. The convective rain rate increases non-linearly as $\overline{T85}$ decreases. This non-linear

Fig. 4. PR reflectivity, $\bar{Z}$, at different altitudes versus $\overline{T85}$ for **a)** convective and **b)** stratiform rain. The symbols denoting the various altitudes are shown as an inset in Panel b.

relationship resembles that derived from radiative transfer theory, where scattering from ice hydrometeors above the freezing level is included (see Wu and Weinman 1984). In the stratiform rain regions, $\overline{T85}$ generally does not go below 200 K, and $\bar{R}$ increases slowly as $\overline{T85}$ decreases.

In Figs. 4a and 4b, we present plots of $\bar{Z}$ at different altitudes in the atmosphere as a function of $\overline{T85}$ using the average *pixel* data given in Tables 1a and 1b. From Fig. 4a, we find that as $\overline{T85}$ increases $\log_{10} \bar{Z}$ decreases almost linearly. In Fig. 4b, where the averages of the *pixel*

data for the stratiform rain are presented, we find again that as $\overline{T85}$ increases $\log_{10} \bar{Z}$ decreases almost linearly. Note, in the stratiform rain areas that the magnitude of $\bar{Z}$ in the reflectivity profiles decreases below the melting layer, which is close to 4 km. This decrease is apparently due to evaporation of rain drops below the freezing level.

In convective rain, the PR and TMI observations shown in Fig. 4a demonstrate that there is a relatively strong relationship between $\overline{T85}$ and the vertical profiles of $\bar{Z}$. The brightness temperatures $\overline{T19}$ and $\overline{T37}$ have a weaker rela-

tionship with the radar reflectivity as shown in Table 1a. From this table, we note that the range in value of $\overline{T85}$, $\overline{T37}$ and $\overline{T19}$ from Row A to Row G is about 117 K, 65 K, and 21 K, respectively. The corresponding range in $\overline{R}$ is about 37 mmhr^{-1} . In addition, the rain rate and Tbs in all three channels vary systematically from Row A to Row G, with no minimum or maximum in between. In stratiform regions, a similar but weaker systematic nature is found.

Based on the averages of the *pixels* presented in Tables 1a and 1b, we note that for a given set of brightness temperatures—i.e., $\overline{T19}$, $\overline{T37}$ and $\overline{T85}$ —one can get widely different rain rates depending on the type of rain that is being observed. An example of this can be found by comparing the data in Row F in Table 1a and Row B in Table 1b. From these rows, we find $\overline{T85}$ is close to 220 K, while $\overline{T37}$ and $\overline{T19}$ are close to 259 K and 276 K, respectively. However, we find near-surface rain rate, $\overline{R}$, deduced from PR for the convective rain is 20.2 mmhr^{-1} , while it is 5.9 mmhr^{-1} for the stratiform rain. Thus, based solely on the brightness temperatures in the 85, 37, and 19 GHz channels, one cannot infer satisfactorily the rain rate unless the type of rain, i.e., convective or stratiform, is known. From this evidence, in order to estimate rain rate from TMI, it is critical to specify the nature of rain from some other information.

From the analysis of the nearly simultaneous TMI and PR observations presented above, we conclude that there is a relationship between them. One such relationship is found for convective rain regions, and another for stratiform rain regions (see Figs. 3a and 3b). In order to vividly show this connection between the TMI and PR data, we have substantially averaged them to minimize the variability introduced by the differences in the scan patterns and viewing geometry of the two instruments, as well as inhomogeneities of the hydrometeors in the atmosphere. In the following section, taking the average properties of the vertical profile of Z given by PR, we are attempting to simulate, using a radiative transfer model, the average brightness temperatures observed by TMI.

3. Radiative transfer simulations

In our radiative transfer model we assume local thermodynamic equilibrium and plane-

parallel stratification, and it includes absorption and multiple scattering processes. Details pertaining to the formulation of the radiative transfer model, including the vertical profiles of the hydrometeors and their extinction properties, and the numerical calculation procedures are elaborated in a NASA Technical Memorandum by Prabhakara et al. (2002). Here, only the essential features of the radiative transfer model and the key results are presented. The important information that is needed in these radiative transfer calculations is related to the vertical profiles of the hydrometeors for the convective and stratiform rain areas.

3.1 General considerations

The basic strategy that we have implemented for both the convective and stratiform models is to simulate the brightness temperatures $T85$, $T37$, and $T19$ for each profile of reflectivity, $\overline{Z}$, given in Tables 1a and 1b. Then, these simulated Tbs are compared with the observed $\overline{T85}$, $\overline{T37}$, and $\overline{T19}$ to establish the physical relationship between the TMI and PR measurements. In our models, the radar reflectivity is given by

$$\overline{Z} = C \int D^6 N(D) dD, \quad (1)$$

where $N(D)$ [m^{-4}] is the number density of particles in a given size range dD [m], and D [m] is the particle diameter. In the case of snow or graupel, the particle diameter is represented by the equivalent melted diameter (see Ulaby et al. 1981). The constant C , which depends on the refractive index, differs between water and ice particles. For water, C has a value of 1.0, while C is 0.19/0.93 for ice. The particle size distribution (PSD) is described using an exponential function, as given below:

$$N(D) = N_0 \exp(-\lambda D). \quad (2)$$

In this equation, N_0 [m^{-4}] and λ [m^{-1}] are the intercept and slope parameters of the PSD. The parameter N_0 is given a value based on observations, such as those of Houze et al. (1979) and Sekhon and Srivastava (1970). Then, in order to match $\overline{Z}$ measured by PR, an appropriate value of λ is determined with an iterative procedure that utilizes Equations 1 and 2. Since the nature of the hydrometeor profile is different in the convective and stratiform regions,

the radiative transfer models applicable to these two regions differ.

3.2 Model and results for convective rain region

In our convective model, the surface temperature over tropical land is set at 302 K, the temperature lapse rate is 6.0 K km^{-1} , and for simplicity the surface emissivity is assumed to be 1.0. This gives a freezing level close to 5.0 km. The water vapor distribution in this atmosphere is taken such that the relative humidity is 80% between the surface and cloud base, and 100% above. The cloud base is set at 1.5 km, and the cloud liquid water content below the freezing level is assumed to increase linearly from 0.5 gm^{-3} to 1.0 gm^{-3} as surface rain rate increases from 2.0 mm hr^{-1} to 50.0 mm hr^{-1} . Cloud liquid water content of such magnitude was reported in hurricanes in a study by Jorgensen (1985).

In the convective region, the PSD at all model levels is inferred using a single value of N_0 . Below the freezing level, all hydrometeors are assumed to be water drops. Above the freezing level, we assume that there is a mixed-phase layer. Analyses of differential reflectivity (Z_{DR}) and linear depolarization ratio (LDR) observations made by polarimetric radars in several other studies (Smith et al. 1999; Bringi et al. 1997; Balakrishnan and Zrnic 1990; and Sax and Keller 1980) lead us to infer a layer of mixed-phase particles, consisting of water and ice, to a substantial height above the freezing level, because of vigorous vertical motions in convective updrafts. The relatively strong reflectivity ($>30 \text{ dBZ}$) observed by PR in convective towers above the freezing level (see Fig. 2a) also supports this inference. The detailed nature of these mixed-phase particles—i.e., the fraction and distribution of water and ice in them—is not considered in this study, because of insufficient information. For this reason, in our convective model, we assume for simplicity that there is a mixed-phase layer of finite depth above the freezing level where water drops and frozen graupel exist together in some proportion. The density of water drops and graupel is respectively 1.0 gm^{-3} and 0.4 gm^{-3} . The fraction of reflectivity due to water drops (f_{wat}) and graupel (f_{ice}) at each level in the mixed-phase layer is given as follows:

$$f_{\text{ice}} = \left(\frac{h - h_F}{H} \right)^{0.25} \quad \text{where } h_F < h < h_T \quad (3)$$

$$f_{\text{wat}} = 1 - f_{\text{ice}}.$$

Here h_F is the height of the freezing level, and h is the height of any given level in the atmosphere. Now, if we denote the height of the top of the mixed-phase layer as h_T , then the thickness H of that layer is given by $H = h_T - h_F$. Equation 3 leads to 100% liquid phase at the freezing level, which nonlinearly approaches 0% as h tends to h_T . Above this mixed-phase layer, we assume all hydrometeors are completely frozen graupel with a density of 0.4 gm^{-3} .

Given N_0 and H , we can define the $N(D)$ for the ice and water particles at each level based on the $\bar{Z}$ profile and Eqs. 1, 2 and 3. With this information, we can calculate the Mie volume absorption and scattering coefficients, k_a and k_s , as well as the phase function, $\mathbf{P}$, as a function of height using the procedure developed by Wiscombe (1980). At each level in the mixed-phase layer, the total volume absorption coefficient is assumed to be the sum of those of ice, water, and gaseous components. Similarly, the total volume scattering coefficient is given by the sum of those of ice and water. Furthermore, the phase function at each level in the mixed-phase layer is assumed to be a combination of the phase functions of ice and water, weighted by their proportions. We may note the properties k_a , k_s and $\mathbf{P}$ depend on the real and imaginary parts of the refractive index of water and ice. For graupel the values of refractive index are computed as a function of the particle density, which depends on the volume fraction of air and ice (Liao and Meneghini 2000). This information allows us to calculate numerically T_{bs} at the top of the atmosphere for a particular viewing angle of the radiometer with the help of the radiative transfer equation and boundary conditions. The procedure described above is utilized to simulate the observed $\overline{T85}$, $\overline{T37}$, and $\overline{T19}$ with the $\bar{Z}$ profile given in each row (see Table 1a) and selected combinations of N_0 and H .

In Figs. 5a and 5b, we have plotted $T85$, simulated with the convective model, as a function of the observed $\overline{T85}$ given in each row of Table 1a. In each figure panel, we show three

curves, which represent T85 simulated using three different values of the depth, H , of the mixed-phase layer above the freezing level (1, 3, and 5 km). Furthermore, in order to show the dependence of these results on N_0 , in Fig. 5a we utilize a N_0 value of $2.0 \times 10^6 \text{ m}^{-4}$, while in Fig. 5b, N_0 is assumed to be $2.0 \times 10^7 \text{ m}^{-4}$. To appreciate the effectiveness of these simulations, we present in each one of these figure panels a line labeled **A**, which corresponds to the observed $\overline{\text{T85}}$.

We find from Figs. 5a and 5b, that as the depth, H , of the mixed-phase layer increases, the curves of simulated T85 tend to warmer values, because of reduced ice scattering. Varying N_0 also has an effect on the simulated T85 values. As the parameter N_0 decreases, the simulated curves increase in temperature with respect to line **A**. In view of the several assumptions pertaining to the nature and vertical distribution of hydrometeors introduced in the model, we can only indicate rough values of N_0 and H that can produce reasonable agreement with observations of $\overline{\text{T85}}$. On this basis, we contend that a value of N_0 between $2.0 \times 10^6 \text{ m}^{-4}$ and $2.0 \times 10^7 \text{ m}^{-4}$, and a value of about 3 km for H , can simulate $\overline{\text{T85}}$ in a crude manner.

From the simulations performed for the convective regions, we find that the 85 GHz brightness temperatures given in Table 1a can be simulated crudely. This is because the 85 GHz channel has a fine footprint and strong extinction. As a result, the response in this channel comes primarily from the layers of the atmosphere above the freezing level. In contrast, because of weaker extinction in the longer wavelength channel at 37 GHz significant part of the signal in this channel comes from layers of the atmosphere below the freezing level, and also from the surface. In addition, inhomogeneity of hydrometeors is more pronounced in the broader footprint of this channel. This can introduce appreciable complexity in the data observed in this channel. In the data of the 19 GHz channel, which has even weaker extinction and a broader footprint compared to the 37 GHz, these problems are compounded further. For these reasons, radiative transfer simulations in these longer wavelength channels for convective regions (see Prabhakara et al. 2000) are not as good as

Fig. 5. For the convective regions, simulated T85 versus $\overline{\text{T85}}$ for the following two values of N_0 : **a)** $2 \times 10^6 \text{ m}^{-4}$ and **b)** $2 \times 10^7 \text{ m}^{-4}$. In each figure, the depth, H , of the mixed-phase layer above the freezing level is indicated for each curve. The solid line **A** in these figures representing $\overline{\text{T85}}$ observations is included for comparison.

those for the 85 GHz channel. However, when the difference, (T19-T37), of the simulations of T19 and T37 is considered, these undesirable effects are substantially reduced. This is demonstrated in Figs. 6a and 6b.

3.3 Model and results for stratiform rain region

In the stratiform model, the surface emissivity and vertical profile of temperature are the same as that in the convective model. The water vapor distribution in this atmosphere is taken such that the relative humidity is 80% between the surface and the freezing level, and 100% above. At altitudes above the freezing level (~ 5.0 km), only ice aggregates (snow

Fig. 6. For the convective regions, simulated $T_{19}-T_{37}$ versus $\overline{T_{19}-T_{37}}$ for the following two values of N_0 : **a)** $2 \times 10^6 \text{ m}^{-4}$ and **b)** $2 \times 10^7 \text{ m}^{-4}$. In each figure, the depth, H , of the mixed-phase layer above the freezing level is indicated for each curve. The solid line **A** in these figures representing $\overline{T_{19}-T_{37}}$ observations is included for comparison.

flakes) are assumed to exist. The density of ice aggregates, ρ_{ice} , is given as follows:

$$\rho_{\text{ice}} = \frac{2.0}{D_{\text{sn}}^2}, \quad (4)$$

where D_{sn} [mm] is the diameter of a spherical particle whose volume is equal to the volume of the ice aggregate. Equation 4 is based on a similar expression developed by Magono and Nakamura (1965) (see also Schols et al. 1999). In Eq. 4, the limiting value of ρ_{ice} is set equal to 0.7 gcm^{-3} , as was done by Schols et al.. For the stratiform simulations, we have developed an empirical model for the intercept parameter, N_0 , of the PSD above the freezing level. This model is based on the observational studies of

Houze et al. (1979) and Sekhon and Srivastava (1970). Houze et al. showed that N_0 increases as the atmospheric temperature decreases, while Sekhon and Srivastava revealed that in the stratiform region N_0 decreases as water content in the atmosphere increases. Based on this information, we have developed an empirical equation to give N_0 as a function of height above the freezing level in the atmosphere as follows:

$$N_0(R, h) = N^* \exp(R^{-0.94} + (h - h_F)), \quad (5)$$

where $h - h_F$ is the height, in kilometers, above the freezing level and R is rain rate. In Equation 5, the argument of the exponential function is forced not to exceed a value of 3.0. Also, the value of N_0 is not allowed to exceed $3 \times 10^8 \text{ m}^{-4}$. This maximum value of N_0 is consistent with the observations of Houze et al. (1979). Below the freezing level, N_0 is given as

$$N_0(R) = N^* \exp(R^{-0.94}). \quad (6)$$

The value of the parameter N^* in Eqs. 5 and 6 is varied over a range between 10^5 m^{-4} and 10^7 m^{-4} to evaluate the impact of this parameter on the upwelling brightness temperature.

In our stratiform model, we have a layer of relatively-large, slowly-falling, low-density, melting snow flakes below the freezing level. The depth of the melting layer is taken to be 500 m. To estimate the refractive index of these melting snow flakes, we have adopted the Maxwell-Garnet mixing model of ellipsoidal ice inclusions randomly distributed in a homogeneous water matrix (Meneghini and Liao 1996 and 2000). In this combination, the water contribution dominates for all meltwater fractions (Bauer et al. 2000). In our stratiform model, it is assumed that these snow flakes melt completely and form raindrops at the base of the melting layer.

The hydrometeor profiles described above for the stratiform rain model are constructed to be consistent with the stratiform profiles of $\bar{Z}$ given in Table 1b. As in the convective model, we compute the Mie volume extinction and scattering coefficients, as well as the phase function, as a function of height for these hydrometeor profiles. This information is used to calculate Tbs at the top of the atmosphere for a particular viewing angle of the radiometer with the help of the radiative transfer equation and

Fig. 7. For the stratiform regions, dashed lines indicate simulated $\overline{T85}$ versus observed $\overline{T85}$ for the following three values of N^* : 10^5 , 10^6 , and 10^7 m^{-4} . For comparison, the solid line **A** representing the observations $\overline{T85}$ are presented in the figure.

boundary conditions. From the description of the hydrometeor profile, we note that N^* is the key parameter for the stratiform region.

In Fig. 7, we are presenting the results obtained from the simulations pertaining to the stratiform observations presented in Table 1b. Three curves (dashed lines) shown in this figure correspond to simulations based on three different values of N^* (10^5 , 10^6 , and 10^7 m^{-4}). In addition, a line labeled **A** represents observed brightness temperatures. The 85 GHz simulations presented in Fig. 7 show that when N^* is between 10^6 and 10^7 m^{-4} , a crude agreement between the observed $\overline{T85}$ and simulated $\overline{T85}$ can be obtained. However, in view of the various limitations, our model fails to simulate the weak range in $\overline{T19-T37}$ in the stratiform region. The difference, $(\overline{T19-T37})$, in the stratiform rain regions has a relatively small range between 9 K and 18 K (see Table 1b). By contrast, we may point out that the range of $\overline{T19-T37}$ for the convective rain is about 13 K to 58 K, which we are able to crudely simulate (see Table 1a).

From the radiative transfer simulations presented above using convective and stratiform models, we conclude it is possible to crudely explain the interrelationship between $\overline{T85}$ observed by the radiometer and $\overline{Z}$ profiles mea-

sured by the radar. However, only the difference $\overline{T19-T37}$ can be explained for the convective region. Mainly, in these simulations, a given reflectivity profile is linked to the observed $\overline{Tbs}$ through the profile of the water content and the prescribed microphysics. The microphysical properties are based on observational studies made by prior investigators using aircraft and radars measurements.

From Table 1a, one finds for convective regions that the average PR reflectivity at all levels in the atmosphere increases systematically as the average near-surface rain rate increases. This systematic relationship explains the connection between $\overline{T85}$ and $\overline{T19-T37}$ with the average near-surface radar rain rate for these regions. For stratiform regions, from Table 1b, one finds that the average PR reflectivity at all levels in the atmosphere does not increase systematically as the average near-surface rain rate increases. This is because of significant evaporation of hydrometeors in the stratiform rain regions (Houze 1997). This introduces some difficulty in relating $\overline{T85}$ and $(\overline{T19-T37})$ with the weak average near-surface radar rain rate in the stratiform regions.

4. Conclusions

Hydrometeor distributions associated with rain are highly variable in space and time. Understanding the impact of hydrometeor profiles on radiative transfer with the help of observations and models is needed in developing rain retrieval methods from microwave radiometer data. For this purpose, in this study we have examined the information given by the PR and TMI onboard the TRMM satellite for relatively pure convective and stratiform rain regions. A simple evaluation of the PR and TMI data shows poor correspondence of the information given by them. This is because of differences in the scan patterns and viewing geometry of the two instruments, as well as inhomogeneities of the hydrometeors in the atmosphere. For this reason, we average the PR and TMI data significantly. The average properties of the PR and TMI data presented in Tables 1a and 1b, based on 20 MCS cases over tropical land, are very helpful to compare the observations from these two instruments. These tables, and Figs. 4a and 4b, show that on average there is a re-

relationship between the PR reflectivity profiles and TMI measured brightness temperatures in 85, 37, and 19 GHz. Also, TMI and PR observations in the tropics show that there are two distinctly different relationships between T85 and PR rain rate in convective and stratiform rain regions. As shown in Fig. 3b, the slope of $\bar{R}$ with respect to $\bar{T85}$ is large for convective regions, and small for stratiform regions. Furthermore, from Tables 1a and 1b, one finds that the radiometer data in 19, 37, and 85 GHz channels on the average have a poor ability to discriminate convective and stratiform rain areas. For this reason, a method to discriminate crudely these rain areas was developed in a study by Prabhakara et al. (2000). That method is based on identification of local minima in the spatial distribution of T85.

In order to explain the relationship between TMI measured brightness temperatures and PR reflectivity profiles (see Figs. 4a and 4b), we have performed radiative transfer calculations as discussed in Section 3. In these calculations, we are able to crudely simulate the 85 GHz brightness temperatures in convective and stratiform regions. The simulated values of T19-T37 for the convective rain region can give similar agreement as in the case of the 85 GHz. This suggests that T19-T37 may be used as additional information to T85 to improve rain retrieval from TMI on land.

Liu and Curry (1998, 1999, and 2000) show from aircraft and satellite passive microwave radiometer data that channels near 150 GHz, which have much stronger extinction than the 85 GHz, can give useful information about ice water content in the upper layers of the troposphere. Because of its higher frequency, for the same size of antenna, a channel near 150 GHz can yield a spatial resolution that is about two times better than the 85 GHz. We find that this high extinction and spatial resolution will be extremely valuable in isolating brightness temperature minima corresponding to convective cells in vigorously developing and decaying thunderstorms. Also, 150 GHz radiation is more opaque to water vapor than that at 85 GHz. This makes a 150 GHz channel useful in the middle and high latitudes, where surface contamination becomes a problem at 85 GHz. From these considerations, we contend that addition of a 150 GHz channel to microwave

radiometers with conical-scanning geometry and with spatially continuous observations will be valuable for the remote sensing of rain.

Understanding the relationship between the measurements made by microwave radiometers and convective and stratiform rain is necessary for the future Global Precipitation Measurement (GPM) mission (E.A. Smith, personal communication). In GPM a constellation of satellites with microwave radiometers are expected to be flown without an accompanying radar. This will also be the case with microwave radiometers onboard the Earth Observing System's Aqua satellite and the Japanese satellite ADEOS-II. Thus, the present study has useful applications for future satellite missions.

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