Precipitation Retrieval over Land and Ocean with the SSM/I: Identification and Characteristics of the Scattering Signal

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ABSTRACT

The subject of this study is the identification of precipitation in warm and cold land and ocean environments from the Defense Meteorological Satellite Program’s (DMSP) Special Sensor Microwave/Imager (SSM/I). The high sensitivity of the SSM/I 85.5 GHz channels to volume scattering by precipitation, especially ice above the freezing level, is the basis for this identification. This ice scattering process causes SSM/I 85.5 GHz brightness temperatures to occasionally fall below 100 K. It is demonstrated that the polarization diversity available at 85.5 GHz from the SSM/I allows discrimination between low brightness temperatures due to surface water bodies versus those due to precipitation. An 85.5 GHz polarization corrected temperature (PCT) is formulated to isolate the precipitation effect. A PCT threshold of 255 K is suggested for the delineation of precipitation. This threshold is shown to be lower than what would generally be expected from nonprecipitating cloud water alone, yet high enough to sense relatively light precipitation rates. Based upon aircraft radiometric measurements compared with radar derived rain rates, as well as model calculations, the corresponding average rain rate threshold is approximately 1–3 mm h⁻¹. The majority of precipitation that falls on the earth exceeds this rate.

Because the 85.5 GHz measurements of oceanic storms are often dominated by scattering due to precipitation above the freezing level, while the 19.35 GHz radiances are dominated by emission due to rain below the freezing level, there is independent information about the gross vertical structure of oceanic precipitation systems from the SSM/I. Apparent differences between storms in formative, mature, and dissipating stages are inferred from the diagnosed amounts of ice versus raindrops, and supported by time lapse GOES imagery. Deviations from the average relationship between 19.35 GHz warming and 85.5 GHz cooling are suggested for use as a diagnostic tool to evaluate lower level rain/upper level ice relative abundances. As an example of this capability, overrunning precipitation shows a horizontal offset between the advancing ice layer and the trailing rain area, consistent with idealized conceptual models of warm frontal precipitation.

Part II of this study will address global screening for the precipitation scattering signal, its statistical characteristics, and the false rain signatures frequently caused by snow cover and cold land.

1. Introduction

The satellite retrieval of precipitation information from passive microwave radiances below about 200 GHz has the fundamental advantage of relative insensitivity to cirrus clouds. Liquid phase clouds are somewhat less transparent. Traditionally, “precipitation” refers to hydrometeors reaching the surface of the earth. Here, depending upon the context, it will refer to either large liquid hydrometeors below the freezing level, or large ice hydrometeors above the freezing level. This

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is because passive microwave precipitation retrievals can be grouped into two categories. The first is emission-based, where liquid precipitation causes brightness temperature \( T_B = eT \) increases over a radiometrically cold (usually ocean) background. The second is scattering-based, where precipitation, especially that above the freezing level, causes brightness temperature decreases over a radiometrically warm (usually land) background. The emission method is the best known, based primarily upon work by Wilheit et al. (1977) at 19.35 GHz, and can be used only over the ocean. The scattering method was elucidated more recently with aircraft observations of the effect at 92 GHz (Wilheit et al. 1982), and with Nimbus 7 Scanning Multichannel Microwave Radiometer (SMMR) 37 GHz data (Spenc-
cer et al. 1983). Different forms of the scattering method can be used over land or ocean. Both emission-based and scattering-based methods were foretold by Buetnner (1963). If one is interested in how much precipitation reaches the surface of the earth, the emission method should give the most direct estimate, while the scattering methods will be more indirect. However, if one is interested in how much precipitation is suspended in the atmosphere, both can be regarded as being good measures of the concentration of precipitation above (scattering) and below (emission) the freezing level. Here we focus on an 85.5 GHz scattering-based method, but use the 19.35-GHz emission-based methods for comparisons in the oceanic rain cases. To retrieve useful precipitation information, it is preferred that the precipitation exhibit a spectral signature sufficiently different from any other geophysical signature.

The previous satellite studies were limited to a highest frequency of 37 GHz, flown on the Nimbus-7 and Seasat SMMR and the Nimbus-6 ESMR. The June 1987 launch of the first Special Sensor Microwave/Imager (SSM/I) on the Defense Meteorological Satellite Program (DMSP) F8 satellite has resulted in a significantly improved precipitation monitoring capability from space. This instrument extends the spaceborne observational frequency capability up to 85.5 GHz (Table 1). Combined with 37, 22.235, and 19.35 GHz radiometers (all dual polarized except for 22.235 GHz), these observations should provide the most sensitive precipitation detection capabilities to date. In addition to the SSM/I’s channel uniqueness, its swath width (Fig. 1) is nearly twice that of the Nimbus-7 SMMR, and the instrument operates continuously, compared to the SMMR’s alternate day duty cycle. Additional improvements include a more direct calibration scheme, with both calibration and earth observations coming through a single feedhorn; synchronous rotation of the feedhorn with the antenna, eliminating the need for polarization rotation corrections; coaligned sampling of all channels’ footprints; and lower noise.

Here we describe and test a method for retrieval of precipitation information over both warm and cold land and ocean environments from dual-polarized 85.5 GHz scattering inferences. The method will be developed from theoretical considerations combined with the observed characteristics of the SSM/I radiances, and will be tested with radar observations over both land and ocean. The method will also be compared to 19.35 GHz oceanic precipitation retrievals. These comparisons will also help assign a threshold value for precipitation assignment. Color-enhanced satellite imagery of rain systems over both land and ocean will be used to illustrate the algorithm.

2. Theoretical considerations

a. Theoretical sensitivity to scattering by precipitation-size ice

The Mie theory (Mie 1908) can be used to quantify the radiation absorption and scattering properties of hydrometeors. The most important precipitation variables affecting brightness temperature measurements at any given wavelength are hydrometeor size, phase, number density, and depth of the layer. If we use the theory to address a Marshall–Palmer size distribution, we find that as we proceed from the lowest SSM/I frequency (19.35 GHz) to the highest (85.5 GHz), we see dramatic increases in the volume scattering ($k_v$) and absorption ($k_a$) coefficients, and single scattering albedo (Fig. 2). It is the high single scatter albedos produced by the ice that are of great importance for the production of microwave radiances that contrast with the precipitation’s surroundings. Ice has much smaller absorption coefficients than water (Fig. 2), leading to high albedos at all SSM/I frequencies.

<table>
<thead>
<tr>
<th>Frequency (GHz)</th>
<th>Nimbus 7 SMMR</th>
<th>F8 SSM/I</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polarization</td>
<td>H, V</td>
<td>H, V (only V at 22.235)</td>
</tr>
<tr>
<td>Radiometer type</td>
<td>Dicke switched</td>
<td>Synchronously rotating</td>
</tr>
<tr>
<td>Swath width</td>
<td>Uncoupled</td>
<td>1394 km</td>
</tr>
<tr>
<td>Scan type</td>
<td>Sinusoidal, zig-zag</td>
<td>Continuous rotation</td>
</tr>
<tr>
<td>Earth incidence</td>
<td>49°</td>
<td>53°</td>
</tr>
<tr>
<td>Spatial resolutions (integrated 3 dB limits, km)</td>
<td>$35 \times 34$ @ 37 GHz</td>
<td>$16 \times 14$ at 85.5 GHz</td>
</tr>
<tr>
<td></td>
<td>$60 \times 56$ @ 21 GHz</td>
<td>$38 \times 30$ at 37 GHz</td>
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<tr>
<td></td>
<td>$68 \times 67$ @ 18 GHz</td>
<td>$60 \times 40$ at 22.235 GHz</td>
</tr>
<tr>
<td></td>
<td>$111 \times 94$ @ 10.7 GHz</td>
<td>$70 \times 45$ 19.35 GHz</td>
</tr>
<tr>
<td>On-orbit delta-T</td>
<td>$1.5° @ 37$ GHz</td>
<td>$0.7° @ 85.5$ GHz</td>
</tr>
<tr>
<td></td>
<td>$1.5° @ 21$ GHz</td>
<td>$0.4° @ 37$ GHz</td>
</tr>
<tr>
<td></td>
<td>$1.2° @ 18$ GHz</td>
<td>$0.8° @ 22.235$ GHz</td>
</tr>
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<td></td>
<td>$0.9° @ 10.7$ GHz</td>
<td>$0.4° @ 19.35$ GHz</td>
</tr>
<tr>
<td></td>
<td>$0.9° @ 6.6$ GHz</td>
<td>$0.4° @ 6.6$ GHz</td>
</tr>
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single scatter albedo approaching unity indicates that any thermal radiation upwelling from below an ice layer that is attenuated by the ice will be scattered out of the radiometer’s field of view, with very little ice-emitted radiation to replace it. If the scattering coefficient is large enough (such as at 85.5 GHz), and if there are no significant sources of downwelling radiation from the cosmic background (only 2.7 K) and the overlying atmosphere to be scattered back toward the spaceborne radiometer, very low brightness temperatures can result.

In radiative transfer calculations, if the single scattering albedo is high, then greater scattering coefficients translate into larger brightness temperature depressions. This is illustrated in Fig. 3, where the relationship between $T_B$ and surface precipitation rate has been modeled over both land and ocean surfaces (Wu and Weinman 1984). This model assumes Marshall–Palmer size distributions of ice water contents that decrease with height above the rain layer in a manner consistent with radar study results. At the lowest rain rates, there is only a shallow layer of ice above the rain which then increases in depth and density as the rain rates increase. From this model we find $T_B$ depressions of tens of degrees (at 18 GHz) to almost two hundred degrees at 85.6 GHz. Because of the small size of many precipitating cells, this change in $T_B$ can occur over distances on the order of 100 m. Thus, volume scattering by precipitation is probably the most striking geophysical passive signature in the microwave spectrum.

It must be emphasized that the assumption of a Marshall–Palmer size distribution for the ice in all
theoretical treatments is suspect. It is the result of convenience, due to the lack of good direct measurements of ice particle size distributions within convective storms. Nevertheless, we can expect that even with more realistic precipitation particle size distributions the strong frequency dependence exhibited in Figs. 2 and 3 will remain. Unfortunately, there are very few in situ measurements of ice particle size and habits in any but the weakest precipitation systems at heights much above the freezing level. Even if complete measurements did exist, radiative transfer modeling for all but the simplest ice particle shapes is very difficult. Therefore, we will rely heavily upon empirical relationships derived from the SSM/I radiances and comparisons with radar data to establish a baseline precipitation diagnosis and retrieval procedure.

Also, while we will emphasize the role of ice above the freezing level in producing 85.5 GHz brightness temperature depressions, it is also true that small depressions (10° to 30°C) are possible with liquid precipitation alone. Theoretical calculations by Savage (1976) indicated that near 100 GHz, $T_b$ below 260 K are easily achievable with rain rates of only a few millimeters per hour. However, due to the lower single scatter albedos for liquid precipitation, a saturation point is quickly approached which is several degrees below the thermometric temperature of the top of the rain layer. Thus, it is the high albedo that ice produces that is necessary to produce very low $T_b$.

b. Polarization correction for water surfaces and the effects of cloud water

Surface water bodies have low emissivities in this portion of the microwave spectrum, causing low brightness temperatures as well. Fortunately, the emissivities are a strong function of polarization for oblique viewing angles. In contrast, it is expected that volume scattering by ice particles should show a much smaller polarization effect (Wu and Weinman 1984), due to their largely random orientation and nonflat habits. Therefore, to help alleviate the ambiguity between water surfaces and precipitation, we can utilize the SSM/I’s polarization diversity. Polarization information has been demonstrated to be of use in separating highly polarized radiances of the ocean from essentially unpolarized radiances from precipitation volume scattering (Weinman and Guetter 1977; Spencer 1986). From an examination of four months (9 July to 31 October 1987) of global SSM/I observations of storms over land, we find that the 85.5-GHz $T_b$ of storm cores are essentially unpolarized (equal amounts of horizontally and vertically polarized radia-

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Fig. 2. Mie volume scattering coefficients (top), volume absorption coefficients (middle), and single scattering albedos (bottom) for a Marshall–Palmer precipitation size distribution of water and ice spheres.
tion). This is true for even the lowest \( T_B \) (72 K) found during this period.

Because the polarization correction for water surfaces is best illustrated with nonprecipitating clouds of varying opacity, it is convenient to address the effects of surface water and nonprecipitating clouds together. This is demonstrated in Fig. 4, where the brightness temperatures of various amounts of nonprecipitating cloud water over the ocean are presented at 19.35, 37, and 85.5 GHz. The radiative transfer calculation assumed a cloud extending from 950 mb to 600 mb within (a) a tropical atmosphere and (b) a standard atmosphere. Both vertically and horizontally polarized \( T_B \) show warming due to clouds, with the effect increasing dramatically with frequency.

Of interest to the current problem is what we will call the polarization corrected temperature

\[
PCT = \frac{\beta T_{B_h} - T_{B_v}}{\beta - 1}
\]

where

\[
\beta = \frac{T_{B_v} - T_{B_h}}{T_{B_h} - T_{B_v}}.
\]

Here, \( T_{B_h} \) and \( T_{B_v} \) refer to the horizontally and vertically polarized cloudfree ocean \( T_B \), respectively, while \( T_{B_h} \) and \( T_{B_v} \) are horizontally and vertically polarized \( T_B \) that are at least partially affected by any combination of clouds and precipitation. \( T_{B_v} \) and \( T_{B_h} \) are the vertically and horizontally polarized \( T_B \), respectively, of the ocean with no overlying atmosphere. The calculation of PCT at any given frequency involves a nearly constant value for \( \beta \) (Fig. 5). Here, \( \beta \) is the ratio of \( T_{B_h} \) warming to \( T_{B_v} \) warming resulting from only atmospheric gaseous absorption (primarily water vapor and oxygen). Model calculations for a standard tropical atmosphere at 85.5 GHz, a 53° view angle, over a calm ocean in thermal equilibrium with the atmosphere (287 K) leads to \( \beta = 0.35 \) and a PCT of 276 K. Similar calculations for a tropical airmass over an ocean with a 302 K surface temperature yields \( \beta = 0.38 \) and a PCT of 288 K. These PCT represent the effective radiating temperature of the lower portion of the troposphere, and correspond to the thermometric temperature near the 80 kPa surface. Note from Fig. 4 that these PCT are warmer than those produced by cloud water.

When the PCT are computed with SSM/I measured \( T_B \) of moist tropical oceanic airmasses (with \( \beta = 0.35 \)) or dry high latitude oceanic airmasses (with \( \beta = 0.38 \)), the PCT are consistently 10° to 15° lower than the model calculations for cloudfree oceanic airmasses. This bias could be the result of wind roughening of the ocean surface, or errors in the model treatment of the sea surface at 85.5 GHz. Because we wish to compute PCT which are physically meaningful (in the range of 275 K to 280 K for most atmospheres), we will use an empirically modified value for \( \beta \). Examination of several days of SSM/I observations of global cloud-free oceanic areas reveals that \( \beta = 0.45 \) gives PCT in the desired range of 275 K to 290 K. Absolute accuracy for \( \beta \) is not as important as keeping it constant in all subsequent calculations, so that a meaningful threshold can be identified for possible precipitation delineation. With \( \beta = 0.45 \), Eq. 1 becomes

\[
PCT = 1.818 T_{B_h} - 0.818 T_{B_v}.
\]

(3)

This is the parameter whose characteristics we will examine in some detail. It should be noted that, if the \( T_B \) are unpolarized, then PCT = \( T_{B_v} \). Also, the impact of any small changes in PCT due to wind roughening anomalies, etc., will approach zero as the difference between \( T_{B_h} \) and \( T_{B_v} \) approaches zero.

Cloud water will generally be distinguishable as 85.5 GHz PCT lower than the background PCT, although its \( T_B \) depression will be much less than that due to the strong volume scattering effects of precipitation. We see in Fig. 6 that cloud water causes brightness temperatures to decrease as the cloud water content or cloud thickness increases. Note that all but the deepest and wettest clouds have 85.5 GHz \( T_B \) above about 250 K. Further increases in cloud water content beyond 0.6 g/cm² cause little additional decrease because the \( T_B \) merely approach the thermometric temperature of the top of the cloud. We will assume that clouds taller
than the 35 kPa surface (8 km) are generally unable to support such a large moisture content without imminent partial conversion to precipitation.

Of critical importance to the present study, thus, is that these cloud-induced depressions are considerably smaller than that which might be expected from all but the lightest precipitation at 85.5 GHz (Fig. 3). When the PCT is calculated for the general case where both cloud and precipitation partially obscure the ocean surface (Fig. 5), we obtain a storm PCT that is lower than what would be expected for the cloud alone (because of volume scattering). Thus, the PCT is useful for determining the extent to which the brightness temperatures are affected by volume scattering, versus by low-emissivity water bodies. Spencer (1986) found that a similar operator applied to Nimbus-7 SMMR measurements resulted in polarization corrected $T_B$ depressions that were well correlated with radar derived rain rates over the Gulf of Mexico coastal waters. In reality, however, there will typically be $T_B$ contributions from both precipitation and cloud water within a rain cloud, so that defining a threshold below which precipitation alone exists and above which cloud water alone exists is not possible. Nevertheless, based upon Fig. 6 we might expect that 85.5 GHz cloud observations below about 250 K to 260 K have a good prob-
ability of being influenced by precipitation. In the following section we will examine a case study of active and dissipating convective activity as viewed by airborne radiometers and a ground-based research radar to help determine a rain rate threshold for this brightness temperature range.

3. Aircraft 92 GHz and CP-2 radar observations of convection

a. The Cooperative Huntsville Meteorological Experiment (COHMEX)

The COHMEX was a 1986 field experiment designed to investigate summertime subtropical continental convection and its effects on the atmosphere (Dodge et al. 1986). As part of COHMEX, NASA high-altitude aircraft (ER-2) flights were periodically made from the Wallops Island Flight Facility at Wallops Island, Virginia to storm targets primarily in northern Alabama and central Tennessee. The ER-2 carried the Advanced Microwave Moisture Sounder (AMMS, a 92 and 183-GHz imager), along with several other instruments concentrating on storm electrical activity and cloud top structure. Several research radars were in operation during COHMEX. The National Center for Atmospheric Research CP-2 radar was designed to provide information on storm structure during the ER-2 flights for later data comparison. This is a dual wavelength (10.68 and 3.2 cm) Doppler radar having polarization diversity and a beamwidth of 1° at both wavelengths. It typically took sector-volume scans of many of the storms that the ER-2 sampled. Best CP-2 polarization measurements were possible within 80 km
of the radar site, allowing discrimination between liquid and frozen precipitation.

b. A COHMEX case study

A small and isolated storm couplet centered 80 km southwest of CP-2 in northwestern Alabama was simultaneously overflown by the ER-2 and probed with the radar at approximately 2155 UTC 11 July 1986. The cold scattering features visible in the resulting AMMS 92 GHz images were easily related to precipitation features measured by the radar, allowing navigation of the aircraft location with about 2 km uncertainty. The ER-2 had accurate heading information from an inertial navigation system (INS) which, together with the AMMS 92 GHz feature matching, was used to isolate the vertical slice through the storm that the aircraft passed over. The CP-2 10-cm reflectivity data for this slice were converted to rain rates by assuming a Marshall–Palmer size distribution consisting of water spheres below 4 km altitude, ice spheres above 8 km, and a linear combination between. The Marshall et al. (1955) relationship between reflectivity and rain rate \( Z = 200R^{1.6} \) was then adjusted based upon the lower reflectivity experienced from ice. This adjustment results in rain rates 2.7 times higher for ice alone than if the reflectivities were assumed to be due to water alone. The resulting vertical distribution of rain rates along the ER-2 flight track (Fig. 7a) reveals a core of active convection near the 30 km mark with rain rates exceeding 60 mm h\(^{-1}\), tilting toward a dissipating anvil having rain rates less than 6 mm h\(^{-1}\) at the surface between the 5 and 20 km marks, and generally less than 1 mm h\(^{-1}\) in thinner portions of the anvil blowing off left of the 10 km mark. The 92 GHz brightness temperature depressions along the ER-2 subtrack roughly match the rain intensity features below the freezing level (Fig. 7b). A scatter plot of the 92 GHz
and radar data (Fig. 8) suggests a good correspondence between brightness-temperature depressions and rain rate. The relationship is noticeably nonlinear, due to the high sensitivity of this wavelength to precipitation size ice. The low $T_B$ outliers near zero rain rate in Fig. 8 are due to the overhang of the strong convective region above an area having virtually no precipitation near the surface (between the 20 and 25 km marks). We expect that this elevated precipitation would eventually fall below the freezing level, which would cause these data outliers to fall closer to the trend exhibited by the rest of the data.
We note from Fig. 8 that a 92 GHz $T_B$ of 255 K in this case corresponds to approximately 3 mm h$^{-1}$. This is not much different from the 2 mm h$^{-1}$ predicted by the Wu and Weinman (1984) modeling study. It is significant that most of the rainfall on the earth probably falls at rates above this, especially in the tropics. For instance, one year of research rain gage data collected at Miami, Florida revealed that about 80% of all precipitation fell at rates greater than 5 mm h$^{-1}$ (Jones and Sims 1978). Because attempts to discriminate progressively lower precipitation rates will result in increasing ambiguities from cloud water, we will tentatively assume a $T_B$ threshold of 255 K for the SSM/I 85.5 GHz delineation of rain areas having average rain rates of about 3 mm h$^{-1}$ or greater.

It is not our purpose to justify these as the best thresholding values of brightness temperature and rain rate to be used with the SSM/I data. Instead, we wish to provide a working basis for the examination of resulting SSM/I patterns of diagnosed precipitation for their characteristics and utility. We hope this will encourage further refinements from future investigations.

4. Precipitation systems observed by the SSM/I

a. Precipitation over land and polarization correction

A squall line over Wisconsin was imaged by the SSM/I on 28 July 1987 (Fig. 9). As in the COHMEX case, the effect of volume scattering by precipitation is seen to be strongly frequency dependent. The 85.5 GHz $T_B$ (Fig. 9a) plummet to below 150 K, but 37-GHz $T_B$ (Fig. 9b) fall to only 230 K, and the 19-GHz $T_B$ (Fig. 9c) dip only a few degrees to about 260 K. The frequency dependence is exaggerated somewhat by the increasing footprint size with decreasing frequency, causing additional spatial averaging.

It is noteworthy that the Great Lakes also have 85.5 GHz $T_B$ as low as the periphery of the storm. However, if we examine a profile of horizontally and vertically polarized 85.5 GHz $T_B$ in a traverse across the storm and Lake Michigan (line AB in Fig. 9), we note that the lake is highly polarized but the storm core is essentially unpolarized (Fig. 10). As discussed in the previous section, correction for polarization effects allows one to correct for the presence of surface water. While it could be argued that water bodies might be avoided with geographical information alone, such a procedure would be time consuming and would not allow for small navigation errors in the satellite data, the presence of transient regions of wet ground, or as will be shown later, the effect of a highly polarized water body shining through a precipitation layer. Here, Eq. 3 was applied to the bipolarized 85.5 GHz $T_B$ and the resulting PCT image (Fig. 9d) shows that any ambiguity between precipitation and water bodies has been eliminated.

Of some interest in Fig. 10 is a small amount of polarization (5° to 10°C) in the stratiform regions of
the storm that does not exist in the cores. Because there was no evidence of a corresponding 37 GHz polarization signature, we can assume that the effect is not due to soil wetting by rain. This result leads us to believe that the SMMR 37 GHz polarization effects reported by Spencer et al. (1983) were probably the result of a calibration bias in one of the channels at low $T_B$, and not due to precipitation scattering. Because the SSM/I calibration strategy has numerous advantages over that used with the SMMR, our confidence in the absolute accuracy of the SSM/I data is much higher. One possible explanation for the 85.5 GHz effect found here is that the stratiform regions might be characterized by dendrites, plates, and columns having flat portions with some nonrandom orientation, thus producing some polarization. The lack of polarization in the cores, on the other hand, might be due to irregularly shaped graupel tumbling in a turbulent environment. This curious feature of the 85.5 GHz data should be examined further for potential insight into storm microphysical structure.

The polarization corrected image in Fig. 9d is color enhanced for all $T_B$ less than or equal to 255 K. It is significant that within 10 min of the SSM/I observation time, both Wausau and Volk Field, Wisconsin National Weather Service Offices reported the start of precipitation at the surface. Both of these stations were as close to the 255 K isotherm as can be determined with...
data having 15 km spatial resolution and small navigation errors.

The GOES infrared image of the Wisconsin storm at the SSM/I observation time (Fig. 9c) shows that a large portion of the thunderstorm cirrus anvil is transparent at 85.5 GHz. About 50% of the anvil area has the same 85.5 GHz $T_B$ as the cloudfree areas surrounding the anvil (272 K to 276 K), indicating that there is no perceptible cooling by the cirrus ice particles. The SSM/I footprint locations for all polarization corrected $T_B$ less than or equal to 255 K are shown overlaid on the IR image (Fig. 9f), with different colors representing different PCT ranges. Note that, for this storm, the infrared 220 K isotherm (black area) corresponds quite closely in area and shape with the 85.5 GHz 255 K footprint locations. The slight cooling that exists closer to the precipitation region (see the 270 K isotherm) could be due to any combination of cold outflow air from the storm cooling the ground, very light precipitation, or cloud water.

Radar reflectivity features from microfilms of the National Weather Service WSR-57 radar PPI (Plan Position Indicator) displays at Neenah, Wisconsin, and Minneapolis, Minnesota, were combined for this squall line. Even though these images show only six reflectivity levels (1 through 6, corresponding to reflectivities of 18, 30, 41, 46, 50, and 57 dBZ, and Marshall–Palmer rain rates of at least 0.5, 2.7, 13.3, 27.3, 48.6, and 133 mm h$^{-1}$, respectively), they are useful for at least delineating precipitation areas. Comparison of these features with the 85.5 GHz 255 K isotherm revealed excellent agreement between radar reflectivity level 1 area shape and coverage and the area contained within the 255 K isotherm (Fig. 11). The only level 1 reflectivity features not delineated by the 255 K isotherm are very small convective elements on the northern end of the squall line, and a weak lobe of low reflectivity on the southern end. Therefore, for this case the 255 K isotherm maps most of the light precipitation (say at least 1–3 mm h$^{-1}$). Even though there are a few radar level 1 reflectivity features missed by the 255 K isotherm, they are still associated with small $T_B$ depressions (to around 265 K). Unfortunately, they are no colder than the land areas just below Lake Superior, and as we

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**Fig. 11.** Comparison between the 85.5 GHz PCT field (left) with the reflectivity patterns (right) observed by the WSR-57 radars at Minneapolis, Minnesota (MSP) and Neenah, Wisconsin (EEW). All times are 1130 UTC 28 July 1987. Both fields are plotted in the radar PPI (plan position indicator) format. The range rings are marked in nautical miles from the radar site. The radar reflectivity levels alternate as 1 (white), 2 (black), 3 (white), etc. The levels 1–6 correspond to minimum reflectivities of 18, 30, 41, 46, 50, and 57 dBZ, respectively, and minimum Marshall–Palmer rain rates of 0.5, 2.7, 13.3, 27.3, 48.6, and 133 mm h$^{-1}$, respectively. Also indicated are the radar beam elevation angles for the radars at the observation time.
shall show later, such $T_b$ are widespread in weather systems producing much cumuliform cloudiness. The colder $T_b$ features embedded in the squall line match most of the higher reflectivity radar features, including several convective cores, and a more uniform stratiform region extending northeastward.  

Next we will briefly review two ocean cases where a similar comparison was made between the 85.5 GHz PCT and radar reflectivity. This will provide a basis for applying the polarization correction technique to precipitation systems in remote oceanic areas.

b. Two oceanic radar-SSM/I comparisons

Simultaneous WSR-57 radar and SSM/I coverage of rain systems off the east coast of the United States are shown in Fig. 12. The first case is southeast of the Carolinas, where deep convection was situated south of a frontal boundary. The SSM/I 85.5 GHz PCT (Fig. 12a) reveals a variety of sizes and intensities of scattering signatures. The size and shape of the corresponding radar level 1 reflectivity field (Fig. 12b) qualitatively matches all features delineated by the 255 K isotherm, with one exception. That is the broad area of level 1 reflectivity (0.5 to 2.7 mm h$^{-1}$) on the southeast side of the precipitation region which has $T_b$ generally in the 255 to 265 K range. As in the Wisconsin case, all of these features together support a 1–3 mm h$^{-1}$ threshold for the 255 K isotherm.

The other case is a nearly circular area of scattering east of New Jersey (c). Very cold $T_b$ are embedded within this storm complex. Again, the simultaneous radar coverage (d) validates the areal distribution of precipitation features with reasonable correspondence between the satellite and radar observed embedded stronger features. The utility of the 255 K isotherm is again evident, although it would be difficult to argue its superiority over the 250 K or 260 K isotherm due to a strong $T_b$ gradient along the edge of the storm.

Both of these oceanic cases demonstrate the potential of utilizing the SSM/I 85.5 GHz data for delineating precipitation as well as for qualitatively indicating its intensity. We will now address a tropical storm over the open ocean where radar data are unavailable.

c. An eastern Pacific tropical storm

An eastern Pacific tropical storm (named “Greg”) was simultaneously imaged by the GOES-East infrared sensors and the SSM/I on the evening of 30 July 1987. Greg had winds estimated at 23 m s$^{-1}$ near the SSM/I observation time. Two major bands of deep convection are seen in the IR image (Fig. 13a) spiraling into a developing eye. Also illustrated are the edges of the SSM/I data swath. The corresponding 85.5 GHz image (Fig. 13b) shows cold areas (green, blue and magenta) embedded in warmer (yellow) regions. These warm (yellow) areas are best explained as low level cumuliform clouds being sensed through the extensive cirrus shield evident in Fig. 13a. The embedded cooler areas (green) we might deduce are precipitation, but we also find that, just as in the Wisconsin case, they are as cold as the oceanic areas distant from the storm. Applying the same polarization correction used in Fig. 9, we find that the water surface/precipitation ambiguity is again eliminated (Fig. 13c). The white areas are warmest and are probably devoid of cloud water. Many slightly darker (somewhat cooler) regions are explainable with cloud water alone. As in the Wisconsin case, all PCT below 255 K are diagnosed as precipitation and are color coded.

If we overlay the locations of the 85.5 GHz diagnosed precipitation on the IR image (Fig. 13d), we see that its areal extent is much smaller than that of the cold IR cloudiness. Time lapse IR imagery reveals that feature “A”, the coldest IR feature, has reached peak development at the SSM/I overpass time. This very cold (as low as 192 K) thunderstorm-produced cirrus layer surrounds embedded cores of 85.5 GHz $T_b$ depressions. These $T_b$ fall as low as 150 K. Just southwest of “A”, feature “B” is a rapidly decaying (warming) anvil, but still very cold (as low as 202 K). Significantly, the 85.5 GHz PCT for “B” showed no perceptible cooling, with values the same as the periphery of Greg (around 280 K). This again demonstrates the ability of the 85.5 GHz radiation to penetrate thick, nonprecipitating anvils with very little attenuation.

Next to “B”, feature “C” is an area of considerable 85.5 GHz scattering. In contrast to “B”, however, it has much warmer IR temperatures, and its time history and spatial characteristics would not have suggested the presence of precipitation. These three, closely spaced features (A, B, and C) demonstrate the independent information contained in the infrared and microwave measurements of precipitation systems.

Since this rain system is over the ocean, we have another independent method for retrieving rain rates to which we can compare the 85.5 GHz scattering measurements. Brightness temperature increases due to precipitation over the ocean at 19.35 GHz are virtually independent of precipitation drop size distribution (Wilheit et al. 1977), making emission-mode radiometry one of the best methods for remotely measuring rain rate. Possibly its greatest limitation is that clouds with high cloud water contents produce some warming which can be misinterpreted as light rain. All 19.35 GHz $T_b$ greater than 200 K have also been plotted over the IR image (Fig. 13e), showing those areas of rain and (to a lesser extent) high cloud water content. We find that the rain-dominated 19.35 GHz large scale structures are very similar to the 85.5 GHz polarization corrected structures. If we average the 85.5 GHz polarization corrected $T_b$ to approximately the same spatial resolution as the 19.35 GHz data (using a moving average of $3 \times 5$ 85.5 GHz footprints, see Fig. 1) we find that there is indeed a good correlation between the two signatures (Fig. 14). Also shown are specific
Fig. 12. As in Fig. 11, but for rain systems observed over the Atlantic Ocean by the Wilmington, North Carolina (ILM) radar at 1109 UTC 29 July 1987 (top); and by the Atlantic City, New Jersey (ACY) radar at 1043 UTC 31 July 1987 (bottom). The arrow in the ILM case shows the largest area out of all three radar cases having level 1 reflectivity (0.5 mm h\(^{-1}\)) that was not delineated by the SSM/I 85.5 GHz PCT of 255 K. Some ground clutter is evident within about eighty nautical miles of ILM.
FIG. 13. (a) GOES-East infrared image (with MB enhancement) of Tropical Storm Greg at approximately 0138 UTC 30 July 1987. The SSM/I swath edges are also shown. (b) SSM/I vertically polarized 85.5 GHz brightness temperature image of Tropical Storm Greg at approximately 0130 UTC 30 July 1987. (c) SSM/I 85.5 GHz PCT image of Greg with all PCT below 255 K color enhanced. (d) SSM/I 85.5 GHz diagnosed precipitation (dots) overlaid on the GOES-East image of Greg. Magenta represents 240 K to 255 K; cyan, 225 K to 240 K; and yellow, 150 K to 225 K. (e) Location of SSM/I horizontally polarized 19.35 GHz brightness temperatures above 200 K (dots) overlaid on the GOES East infrared image of Greg. Magenta represents 200 K to 220 K; cyan, 220 K to 240 K; and yellow, 240 K to 258 K. See text for details.
plane parallel radiative transfer model results for 1) nonprecipitating clouds alone, 2) rain alone, 3) rain capped by a thin layer of ice, and 4) a more sophisticated finite cloud model (which allows radiation to also be scattered out the cloud sides) with a deeper ice layer. At the top of the distribution, we can explain the data with nonprecipitating clouds alone. As we proceed down and to the right in Fig. 14 to colder 85 GHz $T_B$ and warmer 19.35 GHz $T_B$, we can progressively invoke rain, and then ice, to explain the trend of the dual-frequency brightness temperatures. While these model curves do not fully represent the many possible model combinations of cloud water content, rain water content, ice size and number density, and the layer depths of all of these, the trend exhibited by the data are consistent with and most reasonably explained by increasing sizes and number densities of both liquid and ice hydrometeors. From Fig. 14 we find that the 255 K threshold we are testing here corresponds to an average 19.35 GHz $T_B$ of about 220 K. From the 19.35 GHz radiative transfer modeling by Wilheit et al. (1977) and Savage (1976), and taking into account the differences in ocean surface, we find that 220 K (and thus the 85.5 GHz 255 K threshold) corresponds to an average rain rate of about 3 mm h$^{-1}$. This agrees with the value assumed until now.

The most notable outliers in Fig. 14 are very cold 85 GHz $T_B$ with unusually cool 19.35 GHz $T_B$. These are from feature "A" on the northern end of the primary rain band in Fig. 13. Polarization correction of the 19.35 GHz data (not shown) reveals that volume scattering has produced only a few degrees depression. Thus, these cool 19.35 GHz $T_B$ are dominated by the ocean surface shining through the precipitation, not by a precipitation scattering effect (as possibly implied by the proximity of the finite cloud model curve). The most likely explanation for this anomaly is that strong convective updrafts are producing large amounts of precipitation in a deep layer above the freezing level, with as yet relatively little precipitation reaching a low enough altitude to form rain. The extreme coldness and peaking intensity of the IR signature for feature "A" are consistent with this diagnosis. Thus, it is possible that such deviations from the average relationship in Fig. 14 are an indication of the relative abundance of frozen precipitation versus rain. In the following case, we will see a different type of deviation, that indicating a larger proportion of rain relative to the amount of ice.

d. Colder airmasses and a South Atlantic winter cyclone

While we can not say that the $T_B$/rain rate threshold used up to this point is applicable in colder environments, it is instructive to apply it to a winter weather
system if only to uncover obvious differences between summer and winter precipitation systems. Nevertheless, there is some reason to expect about the same 85.5-GHz brightness temperatures in both warm and cold environments for at least light precipitation. Because only a couple of millimeters per hour of rain are needed to virtually obscure any underlying surface at this high frequency, the $T_B$ of light rain should not depend much on whether the freezing level is at 2 km altitude (a cool airmass) or at 5 km altitude (a warm airmass). In either case the source of radiation is dominated by rain in a layer below the freezing level, a layer which has essentially unvarying temperature between airmasses. Of course, differences in the amount of ice above the rain will have a great impact on the resulting radiances and will especially impact their utility for inferring rain rates near the surface. But the fact that light liquid precipitation alone might produce approximately the same $T_B$ in a variety of airmasses suggests the possible use of a geographically constant threshold for at least the delineation of precipitation areas.

The SSM/I imaged a developing winter cyclone over the South Atlantic on 30 July 1987. The GOES-East IR imagery of this disturbance (Fig. 15a) shows poorly organized, mostly cold cloudiness covering the breadth of the SSM/I data swath. The precipitation scattering signatures at 85.5 GHz are not easily distinguished without applying the polarization correction (Fig. 15b). Clouds (yellow) separate moist (green) air in the north from drier (blue) air to the south. Some of the cooler (green) regions are embedded in and adjacent to the clouds, but their source, whether precipitation or ocean, is not obvious. Polarization correction of the 85.5-GHz $T_B$ resolves the ambiguity (Fig. 15c), revealing several bands of diagnosed precipitation. Several are only tens of kilometers wide with rather intense cores, suggesting deep convective activity. The most prominent one is hundreds of kilometers wide, and much more uniform in its intensity, suggesting widespread weaker uplift in stratiform clouds. Based upon cloud motions in time lapse IR imagery, this large feature was found to be on the eastern side of a deep wave in the westerly flow, supporting a warm advection association.

When the 85.5 GHz footprints having PCT values less than 255 K are overlaid on the GOES IR image closest to the SSM/I overpass time (Fig. 15d), we again find that all diagnosed precipitation corresponds to cold cloudiness. But again there is also much cold cloudiness that does not contain the precipitation scattering signature. The 19.35 GHz footprints having significant warming, when overlaid on the IR image (Fig. 15e), show structures somewhat similar to the 85.5 GHz cooling features, but with a north-south offset between the stratiform rain and ice regions. From the 85.5 GHz scattering region outlined in both images, it appears that the southern portion of the ice region at 85.5 GHz could be overrunning precipitation (probably snow) forming at high altitudes and evaporating as it falls (virga). This would explain the lack of a 19.35 GHz rain signature. Farther north, both signatures coexist and reach peak magnitude, suggesting ice above the rain and larger amounts of both. Still farther north, however, only the 19.35 GHz signature is evident, suggesting rain in the absence of significant ice aloft. This analysis is consistent with the classic structure of warm advection precipitation shields when viewed in vertical cross section across the front (e.g. Palmen and Newton 1969).

Another feature of interest is the spot of strong 19.35 GHz warming that has relatively little ice signature (yellow feature "D" in Fig. 15e). This is the reverse of the situation in feature "A" of the tropical storm case, where abundant ice was evident with little rain. The current feature might indicate either a storm being dominated by warm rain processes, or a storm in its dissipation phase with most of its ice already fallen below the freezing level.

The deviations from the average relationships between the 19.35 GHz rain and 85.5 GHz ice signatures in this case, and especially in the tropical storm case (Fig. 14), can be interpreted as special classes of vertical structure in the precipitation. At the most fundamental level, these special classes are i) rain without significant ice above it, or ii) significant ice aloft without rain below. That these differences can be observed passively should be of use in improving our understanding of oceanic precipitation systems in remote regions of the earth.

Based upon the spatial characteristics of this wintertime case, as well as several others not presented, we are hopeful that a constant or nearly constant 85.5 GHz $T_B$ threshold for the delineation of precipitation can be used in both cold and warm environments. However, much more work will be necessary to confidently relate a range of cold brightness temperatures to precipitation rates.

5. Conclusions and recommendations

We have presented both theoretical and observational evidence to support the use of SSM/I 85.5 GHz polarization corrected brightness temperatures (PCTs) for the delineation of precipitation over both land and ocean. A tentative threshold of 255 K has been suggested [when the PCT is calculated with Eq. (3)] for this purpose, with a corresponding rain rate threshold of approximately 1–3 mm h$^{-1}$. These values are based upon the theoretically expected $T_B$ depressions from warm season nonprecipitating cloud alone, the modeled effects of light precipitation, the observed relationship between aircraft radiometer and ground based research radar measurements of a warm season thunderstorm complex, and comparisons between SSM/I radiances and simultaneous radar measurements for several storms. However, the diverse nature of precipitation systems in various stages of their life cycles, and
Fig. 15. As in Fig. 13, except for a frontal wave over the South Atlantic Ocean at approximately 2000 UTC 30 July 1987. Due to the lower $T_b$ encountered with the much drier airmasses for this wintertime case compared to Tropical Storm Greg, the 19.35 GHz color scale (e) is colder than that used in Fig. 14. Magenta, 170 K to 195 K; cyan, 195 K to 220 K; and yellow, 220 K to 250 K. Also shown in all images is the outline of the 85.5 GHz diagnosed major precipitation area, illustrating a southward offset from the 19.35 GHz rain area, consistent with the tilted nature of overrunning precipitation regions.
the uncertainties in radar $Z-R$ relationships and radiative transfer modeling causes us to expect that this average rain rate could easily be in error by a factor of two, and much more in specific cases. It is expected that future work will result in evidence for a somewhat better precipitation threshold to be used with the SSM/I 85.5 GHz channels. Of special interest is the quantification of the 85.5 GHz brightness-temperature-rain rate relationship over its entire range, its variation within the life cycle of a precipitation system, and between precipitation systems in differing atmospheric environments. More measurements of microphysical quantities made in situ, in both cold and warm environments, and in concert with high altitude aircraft radiometric observations, will be necessary to establish a firm basis for quantitative relationships between brightness temperature depressions and precipitation ice water contents aloft and rain rates at the surface.

Despite these uncertainties, there is some reason to expect a uniformly applicable 85.5 GHz $T_B$ threshold for delineation of rain areas having at least some nominal rain rate. Because in weak precipitation systems the layer of light rain just below the freezing level is the primary source of radiation at 85.5 GHz (through large absorption), it might be expected that this layer would produce the same $T_B$ in both warm and cold airmasses. This is because, while the freezing level height changes, the temperature of the rain layer just beneath the freezing level will not change. This hypothesis was qualitatively supported when the 255 K threshold, derived from tropical airmass measurements and modeling, was applied to a winter season cyclone over the South Atlantic Ocean.

There exists a good correlation between the 19.35 GHz $T_B$ warming and 85.5 GHz cooling measured by the SSM/I in tropical oceanic precipitation systems. This suggests a relatively constant rain/ice abundance relationship on the spatial scale of the SSM/I 19.35 GHz channel (60 km). Because the 19.35 and 85.5 GHz radiances can be qualitatively related to the presence and amount of oceanic rain below the freezing level and ice above the freezing level, respectively, deviations from this relationship are interpreted as differences in the gross vertical structure of oceanic precipitation systems. It can be expected that individual convective elements would exhibit a much broader range of rain/ice combinations. The north–south offset between the rain and ice signatures in a large and relatively uniform scattering feature associated with the South Atlantic frontal wave supports its interpretation as a warm front overrunning system.

Above all, this study is intended to provide a basis for further refinement and investigations. Our continuing SSM/I work will deal with the further quantification of the 85.5 GHz PCT — rain rate relationship, global application of the 85.5 GHz algorithm, the statistical characteristics of the diagnosed precipitation radiances, and the effects of cold land, certain geologic features, and snow cover in producing signatures similar to precipitation.

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