SIR-C / X-SAR Observations of Rain Storms

Arthur R. Jameson [1],
Fuk K. Li [2],
Stephen L. Durden [2],
Ziad S. Haddad [2],
Benjamin Holt [2],
Tim Fogarty [2],
Eastwood Im [2],
and
Richard K. Moore [3].

February 13, 1996

Address correspondence to Dr. Z.S. Haddad, 300-227, Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena CA 91109-0899

[1]RJH Scientific, Alexandria, VA
[2]Jet Propulsion Laboratory, California Institute of Technology, Pasadena CA
[3]University of Kansas, Lawrence KS

Abstract

The SIR-C / X-SAR radar observations of rain storms are the first multi-polarization and multi-frequency observations of precipitation from space. In addition to numerous often dramatic images of severe weather systems obtained by forming a synthetic aperture in the usual side-looking attitude, several data takes were performed while the radar antennas were parallel to the ground and the radar beams were pointing at nadir. These opportunities coincided with the passage of the Shuttle over Tropical Cyclone Odille in the southern Indian Ocean during the first flight, and over Typhoon Seth in the Western Pacific during the second flight. The resulting observations, or, more appropriately, the resulting *measurements*, demonstrate for the first time the capability of a spaceborne multi-frequency multi-polarization microwave radar system to quantify precipitation rates, to detect hydrometeor phase, and to classify rain type.

1 Introduction

Improved understanding of the ocean–atmosphere–land heat fluxes requires accurate measurements of rain (Simpson et al, 1988). Indeed, the latent heating/cooling produced by the melting/evaporation/condensation of water is a major driver of global-scale motions in the troposphere. Measuring precipitation profiles obviously requires remote sensing techniques. Microwave radar is one of the important tools for such measurements (see, e.g., Atlas, 1990).

While ground-based radars fulfill a critical role in monitoring severe weather systems that pass within their vicinity, the main advantage of a space-borne radar is that it provides global coverage. Another advantage is that, thanks to its "look-down" geometry, a space-borne radar can measure quite accurately the reflectivity of the hydrometeors within a storm as a function of altitude. In fact, the Tropical Rainfall Measuring Mission (TRMM) satellite, which will be launched in 1997, will include the first quantitative precipitation radar to be flown in space (Okamoto et al, 1988). Its measurements will help determine the nature and the amount of hydrometeors within the different types of storms. Since there are many unknown physical parameters to be estimated in such an inversion problem (including the types of hydrometeors, the size distribution of each type of hydrometeor, and their geometric distribution within the radar beam), a single radar channel can scarcely be sufficient to quantify the 3-dimensional structure of the precipitation with much accuracy. Additional data such as passive radiometer measurements must be used. In anticipation of TRMM's launch, measurements obtained from the SIR-C and X-SAR radars during the Shuttle Radar Laboratory (SRL)'s two missions should prove quite instructive in verifying the detectability and quantitative estimation of precipitation from space, and in testing the feasibility of fusing data from different instruments aboard the same observation platform.

This paper describes the SIR-C / X-SAR radar observations of rain storms, the first multi-polarization and multi-frequency measurements of precipitation from space. These observations were made on SRL-1 in April, 1994, and SRL-2 in October, 1994. A description of the radars can be found in Jordan et al, 1995, and an overview of the two flights can be found in Stofan et al, 1995. The main goals of the rain experiment were to verify the potential for the estimation of rainfall using multiple microwave radar frequencies, to demonstrate the use of polarization in detecting the freezing layer (the radar "bright band"), and to determine the effect of rainfall on the ocean surface backscatter.

Data were collected from two observation geometries: nadir-pointing and side-looking. The nadir observations were obtained by rolling the Shuttle so that the radar antenna beams were normal to the Earth. The radars were then operating effectively as profiling scatterometers. In this geometry, the contamination of the rain echoes by the surface return within the main antenna beam is minimized. The resulting measurements were therefore quite readily amenable to quantitative interpretation. These nadir observations are described in section

2. The side-looking observations were processed using conventional synthetic aperture radar processing. The resulting images and the effects of the rain on the ocean surface are described in section 3.

2 Nadir Observations

When operating in a traditional side-looking observation geometry, the vertical extent of the radar antenna beams essentially integrates the rain returns over the entire depth of the troposphere. Since the change in atmospheric parameters is generally greatest in the vertical dimension, this "integration" in effect obscures much of the important atmospheric stratification. Therefore, in order to minimize this problem, the Shuttle was rolled on a few occasions so that the beams were orthogonal to the surface of the Earth. The challenge of this procedure, however, was the timing of the roll so that there would likely be precipitation in the radar antenna beams some time during the brief period of data collection. By developing decision rationales and procedures over the last few years and, quite frankly, because of good luck, we succeeded in collecting some interesting observations of precipitation in this mode. Before considering these measurements, it is important to mention that precipitating systems fall into two general categories, namely "convective systems" extending vertically over a significant fraction of the troposphere, and "stratiform systems" with negligible vertical air motion, in which the precipitation is largely confined to the lower troposphere (see, e.g., Houze, 1981). The former are associated with more intense precipitation, and, consequently, greater backscattered signals over several kilometers, while the latter are usually characterized by weak signals with the strongest backscatter occurring near the melting level. Consequently, it is a priori feasible to use SIR-C / X-SAR measurements to compute vertical profiles of precipitation in convective systems, but not in stratiform ones. Nevertheless, as we shall see, measurements in stratiform systems can still be quite useful.

Nadir measurements were obtained during one data take (DT 34.55) on SRL-1, and during three data takes (DTs 45.5, 54.0 and 103.0) on SRL-2. During these data takes, measurements were collected with VV and VH polarizations at C- and L-bands, and with VV polarization at X-band. A short pulse length (8.44 μsec) and a narrow bandwidth (10 MHz) were selected for C- and L-bands. The short pulse length minimized the altitude affected by clutter from the surface sidelobes. The bandwidth effect depended on the storm: a large bandwidth would minimize the altitude affected by the surface return; but a narrow bandwidth would minimize the system-noise-to-signal ratio. The choice of the narrow bandwidth implied that for storms with a significant amount of expected total attenuation from the top of the storm to that altitude where the surface clutter takes over, i.e. for storms where the SNR would become uncomfortably low, the SNR would still be maximized. X-SAR was operated with a 40 μsec pulse length and a 9.5 MHz bandwidth. No beam steering was used. The (lowest) pulse repetition frequency, 1.24 KHz, still produced enough pulses to reduce

the fading effects, without introducing any range ambiguities or pulse shadowing effects, and while reducing data storage requirements.

Setting the receiver gain values was less straightforward. On one hand, the mere detection of rain depends only on the SNR, with the accuracy of the quantitative *estimation* of the rain reflectivity improving as the dynamic range of the quantized return increases. On the other hand, the fact that the return from the surface can be as much as 55 dB greater than the return from 0.5 mm h^{-1} rain (Okamoto et al, 1988), and would therefore most probably drive the 8-bit (SIR-C) or 6-bit (X-SAR) quantizers into saturation, places an upper-bound constraint on the receiver gain: as the gain is increased, the increasing echo from the surface would eventually strongly saturate the receiver. It was therefore decided to set the receiver gain so that the return from the top of the strongest expected convective storm was near the top of the dynamic range of the quantizers. This would not avoid saturation by the surface return, but it would still allow one to measure low rain-rates, and to observe the full extent of the storm.

DT 34.55 from SRL-1 produced clear but somewhat disappointingly weak reflectivity profiles from several rain showers within the outer rain band of Tropical Cyclone Odille in the Indian Ocean on April 11, 1994, around 12:42 GMT. The L- and C-band radar operation parameters were incorrectly set during DT 45.5 from SRL-2. Fortunately, DTs 54.0 and 103.0 from SRL-2 produced data with excellent SNR. Both were performed over Typhoon Seth in the Western Pacific, DT 54.0 on October 3, 1994, around 17:44 GMT, and DT 103.0 on October 6, 1994, around 18:14 GMT. Both contain localized echoes from several convective cells as well as extended returns from stratiform rain. The remainder of this section will concentrate on describing one stratiform and two convective cases.

Figure 1a shows the GMS IR image from October 6 at 17:32 GMT, overlaid with the Shuttle track during data take 103.0 . The black cross towards the bottom left of the track marks the location of a convective cell that produced the reflectivities shown in figure 2. Since the GMS image preceded the data take by about half an hour, the brightness temperatures are expected to have changed somewhat by the time of the Shuttle pass. Figure 2 shows the measured VV effective radar reflectivity factors at the three radar frequencies. No surface clutter subtraction was performed on the SIR-C reflectivities because the surface return was saturating the receiver too strongly. However, we were able to subtract the clear-air surface return from the X-SAR data, although some residual contamination was unavoidable. The discrepancy between the C- and X-band reflectivities, on one hand, and the L-band reflectivity, on the other, is due largely to the difference in the along-track antenna beam widths: while the 3-dB beam width is 0.25° at C-band and a comparable 0.15° at X-band, it is 1.1° at L-band. Thus, the obviously significant along-track variation of the intensity explains the large-scale difference between X/C- and L-bands. The along-track variation of the intensity also strongly suggests that the rain was very non-uniform across track too. While a gain corresponding to the nominal $5° - 6°$ across-track 3-dB beam widths was assumed to produce these images, it is much more likely that the rain echo was produced

by precipitation occupying only a fraction of the actual antenna beam. In fact, since the along-track dimension of the "pockets" producing the strongest rain reflectivities manifestly did not exceed 2 km, one can reasonably assume that the same must have been true of the across-track dimension. Since this would correspond to a 0.5° beam width, one can therefore conclude that the measured reflectivities are very likely about $10\log_{10}(5/0.5) = 10$ dB too low because of this partial beam-filling effect.

While no significant cross-polarized return was measurable in this data segment, the return from the second convective cell marked by the black cross on the upper portion of the track in figure 1a did produce a distinct VH return at C-band. Figure 3 shows the C-VV, L-VV and C-VH reflectivities. The C-VH reflectivities were plotted using a specific color scale to enhance what appears to be the return from melting ice particles. Leakage from the V channel can be ruled out since the stronger VV returns from the lower altitudes do not appear in the VH data. Indeed, similar signatures were observed routinely during the Convection and Precipitation/Electrification (CaPE) project in Florida (see, e.g., Jameson et al, 1996), and were found to be due to the melting of denser ice particles.

Figure 1b shows the GMS IR image from October 6 at 17:32 GMT, overlaid with the Shuttle track during data take 54.0 . The black cross about two-thirds of the way up along the track marks the location of the region of stratiform precipitation within the outer edges of Typhoon Seth that produced the reflectivities shown in figure 4. The four panels in that figure show the measured VV reflectivities at the three frequencies as well as the C-VH reflectivity. As before, no surface-subtraction was performed except for the X-band data (where subtracting the clear-air surface return is necessary because of the long illuminating pulse).

There are several interesting features in figure 4. In all panels, the appearance of the "melting layer" at an altitude of about 5 km is striking. While this observation is not new (the so-called "bright band" has been studied over the last 50 years), these observations are not only the first reports from a spaceborne radar, but they are the first collected simultaneously at three frequencies and at two polarizations. The rapid increase in signal intensity responsible for the banded appearance is due largely to the change in the index of refraction as ice begins to melt (see, e.g., Austin and Bemis, 1950; Browne and Robinson, 1952; Newell and Geotis, 1955). The magnitude of this effect, however, is governed not only by the transition to a liquid having a much greater index of refraction, but also by other factors such as the kind of icy hydrometeor, any growth (accumulation of mass) of ice just before melting, and changes in fall speed as melting proceeds. Specifically, icy hydrometeors range in densities from 0.05 to 0.9 g cm^{-3}. Particles having lower densities are usually ice crystals and snowflakes, while hydrometeors having higher densities are the more compact ice particles such as hail and graupel. These differences are important with respect to the signals because, for example, snow and crystals usually fall at a rate of around 1 m s^{-1} while the denser particles can reach terminal fall speeds of several meters per second. Consequently, snow and ice can have a substantial increase in mass (and therefore, radar

reflectivity) just before melting by the processes of vapor deposition and the accretion of supercooled (subfreezing) cloud droplets. Moreover, snow and ice crystals tend to clump as they fall so that they can become low density aggregates having quite large dimensions of one to several centimeters. On the other hand, the relative change in mass of the higher density particles is usually considerably smaller because their substantial terminal velocity assures a rapid passage through the thin layers of supercooled cloud. In addition, these hydrometeors rarely aggregate.

Furthermore, the melting processes of these two types of hydrometeors are different. The denser particles tend to form a film of surface liquid with melting of the ice continuing inside this liquid shell. A horizontally-oriented ring of liquid water can form around the middle of the hydrometeor before melting is completed (see, e.g., Chong and Chen, 1974). Snow, on the other hand, begins by first wetting the outer, most exposed crystals so that the snow at first acts like a clump of randomly oriented dipoles (e.g., see the discussion in Jameson, 1989). As melting proceeds, this three-dimensional asymmetric structure collapses into a more two-dimensional pancake of loosely bound water drops which finally breaks apart once the icy structure loses its integrity (Fujiyoshi, 1986). Because the raindrops fall with their largest dimensions oriented horizontally, they generate very little cross-polarized return unless they are several mm in diameter (Jameson and Durden, 1996), an infrequent occurrence in this type of rain. Furthermore, whereas the terminal fall speed of the denser ice may show little to modest increases in fall speed, the drops from the melted snow fall considerably faster than the original flakes. As a consequence, between these two types of hydrometeors, differences appear in the signals associated with the melting layer. In particular, less dense particles tend to form very shallow (in depth), well-defined radar bright bands (as in figure 4) because the melting occurs in a thin region as a consequence of the slow fallspeed of the snow. Significant cross-polarized signals occur at the melting level where the wet crystals act like randomly oriented wet dipoles. Once the melting is completed, however, the cross-polarized signals disappear with the formation of the raindrops. Signals from the melting of denser hydrometeors (as illustrated in figure 3), on the other hand, tend to be weaker in part because the melting occurs over much greater depths (because of the larger fallspeed and different melting process), and the cross-polarized signals tend to be much reduced because, unlike snow, these particles usually lack significant three dimensional asymmetry during melting.

So what do the structures in figure 5 tell us? Beginning with the C-band vertical polarization panel at the top, a thin melting layer is obvious. There is a hint of precipitation above the melting layer, while shafts of rain are clearly evident below the layer. In the cross-polarized channel at C-band (C-VH), signals are evident with little if any signal above or below the melting layer. This is precisely the scenario anticipated for melting snow and ice crystals, and we can conclude with good confidence that the 0°C isotherm lies near 5 km above the ocean surface, a common height above warm tropical oceans. In addition, as one might expect because of the higher frequency, it appears that the X-band is somewhat more sensitive than the C-band to the presence of ice above 5 km, yielding clear evidence of a band

of snow descending from much greater heights. The sudden increase in the X-band signal below the melting level may, in part, also be the result of depolarization by the precipitation in the melting layer at this higher frequency.

While the increased sensitivity at X-band may not be too surprising, it is rather amazing to see how readily the melting layer is detected at L-band. Since L-band is not particularly sensitive to liquid water (raindrops), such a strong signature at the melting layer must not only be a response to the change in the index of refraction, but must also be due to the presence of very large particles of several centimeter diameter. Very few hydrometeors besides hail reach these dimensions. Yet it is unlikely that these observations involve hail which can only come from highly convective storms. Instead we conclude that the particles are very large aggregates of moist, melting snow. In principle more quantitative estimates of the particle sizes should be possible by considering the ratios of these measurements at different frequencies. However, the beam dimensions at the different wavelengths are so disparate that such studies could be very misleading. Nevertheless, it is reasonable to conclude that we are not only looking at melting ice crystals but also at very large melting snowflakes, and that the melting level is at 5 km.

Although independent confirmation of these observations during the experiment was not possible, they are in agreement with similar observations in different settings using different radars as indicated in figure 5. A profile through the melting level in figure 4 is illustrated in figure 5a. The structures are quite similar to those observed using the NASA/JPL ARMAR airborne radar that flew over a melting layer associated with a forming tropical cyclone in the tropical Western Pacific in 1993, shown in figure 5b. Note the similar structures in both the linear depolarization ratio [defined by LDR $= 10 \log_{10}(Z_X/Z_C)$, where Z_X and Z_C are the radar reflectivity factors at cross- and co-polarization, respectively] and the reflectivity factor, Z. Finally, ground-based NCAR/CP-2 vertically-pointing radar measurements in stratiform precipitation in Florida are shown in figure 5c. While exhibiting similar features, they are displaced downward by about 1 km. Particularly noteworthy is the increase in the downward fallspeed of the scatterers as a consequence of melting. In fact, one can estimate the mass weighted mean size of the drops assuming that the vertical air motion in this kind of situation is negligible compared to the fallspeed of the hydrometeor (which is often the case in stratified systems after melting has occurred). Under this assumption, a fallspeed of 7 m s^{-1} implies a mass weighted drop diameter of around 2.2 mm (Jameson, 1993), a very reasonable value. Thus, when such structures exist, spaceborne observations of these features can be used to derive reasonable estimates of the height of the melting level and even reasonable estimates of the size of raindrops under the right conditions. This is important, for example, because when attempting to derive quantitative estimates of rainfall from radiometric and radar measurements, it is important to know something about the size of the particles and the depth of the region likely to contain rain. For radiometers this is important because rain is a strong emitter of microwave radiation, while for radars, it is a strong absorber of microwaves at higher frequencies. Both processes depend on the size of

the raindrops and the extent of the rain.

As to the measured reflectivities from the convective cells, they can be used to estimate quantitatively the amount of precipitation as a function of altitude. Consider the case of figure 2, where the absence of any detectable echo in the C-VH channel allows one to reasonably assume that the hydrometeors consisted mostly of liquid water. Let $N_r(D)dD$ be the number of drops of diameter D mm per cubic meter at range r: thus, $N_r(D)$ is the drop size distribution (DSD) at range r. Analyses of disdrometer data (Ulbrich, 1983; Haddad et al, 1996(c)) show that the DSD can be assumed to have the form $N_r(D) = N_0(r)D^{\mu(r)}e^{-\Lambda(r)D}$, where the variables N_0, μ and Λ can be parametrized by

$$\mu(r) = \frac{e^{0.034R(r)^{0.74}}}{s'^2 D'^{0.4}} - 4 \tag{1}$$

$$\Lambda(r) = \frac{e^{0.034R(r)^{0.74}}}{s'^2 D'^{1.4} R(r)^{0.155}} \tag{2}$$

$$N_0(r) = 55\frac{\Lambda^{\mu+4}}{\Gamma(\mu+4)\left(1-(1+0.53/\Lambda)^{-\mu-4}\right)}R(r) \tag{3}$$

in which $R(r)$ is the rainrate at range r, and s' and D' are statistically independent DSD shape parameters which, empirical analyses show, can be assumed constant to first order. Given the measured C-VV and X-VV reflectivity profiles, one would expect to be able to retrieve the rainrate profile $R(r)$ and the values of s' and D' that are most likely to have given rise to the observed reflectivities. If the retrieval problem were deterministic, it would seem over-determined, since one must calculate from two given (reflectivity) profiles one (single) rainrate profile along with the value of each of the two (constant) parameters s' and D'. However, the retrieval problem in this case is stochastic rather than deterministic, due to the fading-, system- and surface-clutter noise contributions, and to the unknown partial beam-filling effect. As was shown in Haddad et al, 1996(a,b), one can still apply stochastic filtering techniques to estimate the values of the variables involved, given a forward equation specifying the relation between the measured reflectivities and the DSD variables. An algorithm similar to the one described in Haddad et al, 1996(b), was used to estimate the rainrate profile through the center of the cell shown in figure 2. The measured reflectivity profiles are shown in figure 6. For simplicity, we assumed that only a 0.5° portion across-track of the radar beams was filled with rain, thus adding a correction of about $10\log_{10}(5/0.5) = 10$ dB to the measured reflectivities in figure 6 to account for the partial beam-filling problem. Figure 7a shows the resulting estimated rainrate profile, along with 1-σ error bars which our estimation approach produces. Note that the r.m.s. uncertainty in the rainrate estimate is relatively large at altitudes below about 6 km, corresponding to the added ambiguity in the retrieval due to the extent of the surface clutter at X-band. Once the estimates of R, s' and D' are found, the DSD parameters μ, Λ and N_0 can be obtained from (1)–(3). For example, figure 7b shows the estimated profile of the DSD parameter Λ. In this case, the estimated expected values $(D', s') = (1.1, 0.35)$ imply that the mass-weighted mean drop diameters were average-to-small with a relatively small spread about the mean (Haddad et

al, 1996(c)). It also turned out that, while the partial beam-filling correction does change the rainrate estimates, it does not alter either the shape of the profile or the estimated expected values of s' or D' significantly. Details of the approach will appear elsewhere.

While it is most interesting to study the echoes due to precipitation, the nadir surface returns can be useful to estimate the total attenuation through the rain. Unfortunately, severe receiver saturation made it impossible to estimate the echo power or to analyze its statistics.

3 Side-Looking Observations

As part of the experiment, side-looking data over rain were acquired on data takes 13.4, 46.3, 66.6, 72.9, 140.4, 147.8, and 149.1 on SRL-1. On SRL-2, side-looking data over rain were acquired on data takes 13.4, 77.5, 114.0, 114.6, 119.0, 141.5, and 157.5. Images from these data takes show large modulations of the ocean σ^o in areas where there appears to be rain storms. Because of the side-looking geometry, quantitative interpretation is challenging. Previous side-looking SAR imagery of rain over the ocean has been interpreted in terms of the effects of wind and rain on the ocean surface (see, e.g., Atlas, 1994(a,b)). These analyses, however, have dealt only with single polarization and single frequency data. The multiple frequencies and polarizations used in the SIR-C/XSAR data provide an opportunity to improve the interpretation of the data.

DT 13.4 from SRL-1 is of particular interest because it captured an extensive rain area using mode "16X" (Jordan et al, 1995), giving the full scattering matrix at L- and C-bands. For our analysis the standard SIR-C and X-SAR calibrated image products were used. Figure 8 shows an example of three-frequency VV polarization imagery from this data take. The incidence angle for these data is $\simeq 50^o$. Figure 9 shows a contrast enhanced version of the L-band image with several dark areas noted A-G. For these areas, σ^o is near -40 dB, while the surrounding areas have σ^o near -20 dB. This variability in L-band backscatter is likely to be entirely controlled by ocean σ^o variations due to windspeed (Durden and Vesecky, 1985) and rain (Tsimplis, 1992). Attenuation and scattering by rain at L-band are likely to be small. The L-band attenuation in rain is much less than 1 dB, even for rainrates well over 100 mm/h (Olsen et al, 1978). Also, rain backscatter is expected to be small relative to the return from the ocean, except for a very smooth ocean. At C- and X-bands, interpretation becomes more complicated, since both attenuation and scattering in rain could be significant. Areas A-E, which are darker than the surrounding areas at L-band, are brighter than the surrounding areas at C- and X-bands. These areas have C- and X-band σ^o up to -10 dB in contrast to -40 dB at L-band, while the surrounding areas have σ^o near -20 dB, similar to L-band. The same phenomenon can be see in the area just to the left of F and down to E. This type of increase in brightness with frequency is a common phenomenon, appearing not only in

figure 8 but in many other data takes as well. The most likely explanation is that these areas were being subjected to moderate to heavy precipitation. At C- and X-bands backscatter from the precipitation is significant, increasing backscatter from these areas relative to the surrounding areas. At L-band backscatter from precipitation is not significant. Furthermore the ocean may be smooth due to damping of waves by heavy rain, making backscatter lower than the surrounding area. An explanation based only on ocean scattering would require that the surface be smooth near the L-band Bragg wavelength (16 cm) but very rough near the C- and X-band Bragg wavelengths (4 and 2 cm, respectively). This does not seem to agree with what is currently understood about the effect of rain and wind on the ocean (Phillips, 1977, Tsimplis, 1992, Poon et al, 1992). While A-E become bright as frequency increases, F and G remain dark at C- and X-bands, suggesting that the rainrate in them is low and that the ocean is smooth. In this case, the smooth ocean is probably related to low winds, since the rainrate would probably be too low to damp waves. The area between E and G exhibits decreasing backscatter with increasing frequency. The backscatter cross section drops from -27 dB at L-band to -32 dB at X-band, possibly due to increasing attenuation in precipitation.

Polarization observations provide additional information. The HH to VV ratio (called ZDR in the radar meteorology literature) should be strongly negative for ocean scattering at 50^o incidence (Durden and Vesecky 1985). Over much of the image, including the areas near A-G, HH is around 5 dB lower than VV at C-band; i.e. ZDR $= -5$ dB. However, within A-G, ZDR is generally larger, reaching a maximum of -2 dB. From both theory and observations (Jameson, 1989), scattering from rain generally has positive ZDR, while scattering from ice hydrometeors can be either slightly positive or negative. ZDR for two-scale scattering from the ocean surface would also increase with increasing windspeed, since HH and VV scattering have different windspeed dependences (Durden and Vesecky, 1985). However, an increase in ZDR of 3 dB might require an order of magnitude increase in windpseed. As explained below, ZDR can also be increased by the presence of direct-reflected scattering, which involves scattering from both rain and ocean. At L-band ZDR is consistently around -9 dB, both in A-G and throughout the image, confirming that the scattering is primarily from the ocean. A-E in figure 9 generally have high C-band LDR (up to -10 dB). This level of LDR should be above the SIR-C antenna cross-polarization isolation level (Jordan et al, 1995). Assuming that this is the case, the high LDR is most likely the result of backscatter from melting ice particles. Propagation through precipitation could also contribute to depolarizing the signal, as could ocean scattering. The C-band HH-VV phase difference is low throughout much of the image. However, it is large ($> 100^o$) in many spots in the top portion of the image and particularly in the "hole" regions F and G. At L-band the HH-VV phase reaches 40^o in G and is near zero elsewhere. As noted above, F and G have low backscatter at all three frequencies, indicating that the rainrate is low and the ocean relatively smooth. A non-zero HH-VV phase can be caused by propagation through precipitation. However, in this case, the probable path length and rainrate seem much too low to explain such large phase differences. A more likely cause of the large phase difference in some areas is the

presence of direct-reflected scattering, discussed in detail below.

In addition to data takes over tropical oceans, SRL-1 serendipitously captured several rain cells over the Amazon rain-forest. These data are particularly interesting because while interpretation of SAR imagery of rain over the ocean is complicated by the possibility of backscatter and attenuation in rain, as well as the variability of the ocean σ^o itself, images of rain over the Amazon do not suffer from any significant surface σ^o variability (assuming that rain does not strongly affect backscatter from trees). Figure 10 shows three-frequency imagery of rain cells over the Amazon during DT 103.6 from SRL-1. At X-band the rain cells show up as dark areas, with a slight brightening at the leading edge of each area. At C-band, no brightening is visible, but some dark areas can be seen in the same location as the dark areas at X-band. At L-band neither bright nor dark areas are visible. At all three frequencies the transmitted wave is both scattered and attenuated by the rain. However, at L-band, both effects are small. At C-band the attenuation is large enough to reduce the observed σ^o by the two-way loss through the rain. The rain backscatter is small relative to the tree-top backscatter and is not seen. At X-band the backscattering from rain (-7 dB) is large enough that it can be seen above the cross section of the tree tops (-9 dB). However, as range increases, the signal backscattered from both surface and rain is attenuated strongly. These data are consistent with the rain scattering explanation discussed above for the bright areas in the C- and X-band ocean images (figure 8).

To interpret these observations quantitatively, we consider scattering from a random medium over a reflecting surface. As discussed in Chauhan et al, 1991, for example, scattering in this situation consists of the components illustrated in figure 11: direct backscatter from the random medium, backscatter from the surface, direct-reflected interaction of the surface with the random medium, and backscatter from the random medium, reflected by the surface. The total ocean σ^o is given by

$$\sigma^o = \sigma^o_s + \sigma^o_d + \sigma^o_{dr} + \sigma^o_r \tag{4}$$

where the subscript s denotes surface, d denotes direct from rain, dr denotes direct-reflected, and r denotes reflected. Note that each term includes both scattering by the surface and/or random medium and also attenuation due to the intervening random medium. As mentioned above, the dr term can provide a large phase shift between HH and VV. In the problem at hand, the random medium represents rain, while the surface is the ocean. This model has been used extensively in modeling SAR imagery of vegetation, and has typically been implemented using the Distorted Born Approximation (DBA) and assuming that the surface is only slightly rough, so that the dr and r terms use only the specular or coherent component from the surface (Lang and Sidhu, 1983; Chauhan et al, 1991; Durden et al, 1995). The DBA assumes that the scatterers in the medium "see" the attenuated field. This assumption is also used in the radar equation of radar meteorology and should apply to our case. The assumption of a slightly rough surface, however, cannot be justified for the ocean at the frequencies used here. Hence, both the dr and r terms will be much more complicated than

in the vegetation case, requiring integration over the scattering pattern of the surface. This has, in fact, been carried out previously for the r term in the case of radar observations of rain over the ocean at nadir and turns out to be quite complicated (Meneghini and Atlas, 1986). Their work attempted to use the r term, also called mirror image return, to profile rain. Of special interest in equation (4) is the dr term, in which the transmitted wave forward-reflects from the surface and then reflects from rain back to the radar. This mechanism is analogous to a dihedral corner reflector, which produces a 180-degree phase shift between HH and VV return. Ulaby et al (1987) showed that it could explain large phase differences in SAR observations. Durden et al (1989) used it to explain the large HH-VV phase difference occuring in vegetation over flooded ground. It was also significant in scattering from flooded rice fields, again producing a rather large HH-VV phase (Durden et al, 1995). This mechanism is the likely cause of the large HH-VV phase differences observed in some areas, as noted above. Because the Fresnel coefficient for forward scatter from the ocean is larger for H than it is for V, the presence of this mechanism typically increases ZDR, and may have contributed to the ZDR observations in figure 8.

For the case of backscattering from rain over a rainforest, the model can be simplified to include only the surface and direct backscatter terms. A simple model of the backscatter from a rectangular rain cell over the rain-forest was thus constructed. The model cell is 4.5 km in altitude and 4 km wide (cross-track direction) and has a rainrate of 40 mm/hr. Figure 12 plots the observed and modeled σ^o versus range. The model duplicates the observation quite well, supporting our interpretations. Rain cells such as these can be found in many of the SIR-C/X-SAR data takes over the Amazon rainforest.

4 Conclusions

In summary, the SIR-C / X-SAR mission's data over rain storms have successfully produced the first multi-polarization and multi-frequency measurements of precipitation from space. Rainfall was clearly detectable even at L-band. We were able to estimate rainrates from the C- and X-band nadir returns. Melting hydrometeors were clearly detectable in the cross-polarized C-band echoes from stratiform as well as convective systems, much as one would expect given past ground and air-borne measurements.

The data also confirm previous observations that rain and wind effects can produce large variations in $30° - 50°$-incidence-angle sea surface backscatter. At nadir, however, the ocean surface echo could not be estimated due to severe receiver saturation.

Finally, several areas with significantly non-zero HH-VV phase differences and apparently lighter rainfall were observed in the ocean images.

5 Acknowledgements

We wish to thank Torben Nielsen, Wenxia Yu and Winston Dang at the University of Hawaii, and George Dugdale, Michael Saunby and Neil Ellis with the TAMSAT program at the University of Reading for their invaluable help in obtaining GOES, GMS and METEOSAT data during the mission. We also thank Annette DeCharon and Jeff Levison for their help in planning the radar measurements and obtaining the data, and Rudolf Winter from DLR for graciously sharing his data. This work was performed at the Jet Propulsion Laboratory, California Institute of Technology, under contract with the National Aeronautics and Space Administration.

6 References

Atlas, D.A., editor (1990), *Radar in Meteorology: Battan Memorial and 40th Anniversary Radar Meteorology Conference*, American Meteorological Society, Boston.

Atlas, D.A. (1994(a)), Footprints of storm on the sea: A view from synthetic aperture radar, *J. Geophys. Res.* 99: 7961-7969.

Atlas, D.A. (1994(b)), Origin of storm footprints on the sea by synthetic aperture radar, *Science* 266: 1364-1366.

Austin, P.M., and Bemis, A.C. (1950), A quantitative study of the "bright band" in radar precipitation echoes, *J. Meteor.* 7: 145-151.

Browne, I.C., and Robinson, N.P. (1952), Cross polarization of the radar melting band, *Nature* 170: 1078-1079.

Chauhan, N.S., Lang, R.H., and Ranson, J.K. (1991), Radar modeling of a boreal forest, *IEEE Trans. Geosci. Remote Sensing* 29: 627-638.

Chong, S.I., and Chen, C.S. (1974), Water shells on ice pellets and hailstones, *J. Atmos. Sci.* 31: 1384-1391.

Durden, S.L., and Vesecky, J.F. (1985), A physical radar cross section model for a wind-driven sea with swell, *IEEE J. Oceanic Eng.* 10: 445-451.

Durden, S.L., van Zyl, J.J., and Zebker, H.A. (1989), Modeling and observation of the radar polarization signature of forested areas, *IEEE Trans. Geosci. Remote Sensing* 27: 290-301.

Durden, S.L., Morrissey, L.A., and Livingston, G.P. (1995), Microwave backscatter and attenuation dependence on leaf area index for flooded rice fields, *IEEE Trans. Geosci. Remote Sensing* 33: 807-810.

Fujiyoshi, Y. (1986), Melting snowflakes, *J. Atmos. Sci.* 43: 307-311.

Haddad, Z.S., Im, E., and Durden, S.L. (1996(a)), Optimal estimation of rainrate profiles from single frequency radar echoes, *J. Appl. Meteor.*, 35: 214-228.

Haddad, Z.S., Im, E., Durden, S.L., and Hensley, S. (1996(b)), Stochastic filtering of rain profiles using radar, surface-referenced radar, or combined radar-radiometer measurements, *J. Appl. Meteor.*, 35: 229-242.

Haddad, Z.S., Durden, S.L., and Im, E. (1996(c)), Parametrizing the raindrop size distribution, *J. Appl. Meteor.* 35: 3-13.

Houze, R.A., Jr. (1981), Structures of atmospheric precipitation systems: a global survey, *Radio Sci.* 16: 671-689.

Jameson, A.R. (1989), The interpretation and meteorological application of radar backscatter amplitude ratios at linear polarizations, *J. Atmos. Oceanic Technol.* 6: 908-919.

Jameson, A.R. (1993), The meteorological parametrization of specific attenuation and polarization differential phase shift in rain, *J. Appl. Meteor.* 32: 1741-1750.

Jameson, A.R., and Durden, S.L. (1996), A possible origin of linear depolarization observed at vertical incidence in rain, *J. Appl. Meteor.* 35: 271-277.

Jameson, A.R., Murphy, M.J., and Krider, E.P. (1996), Multiple-parameter radar observations of isolated Florida thunderstorms during the onset of electrification, *J. Appl. Meteor.* 35, in press.

Jordan, R.L., Honeycutt, B.L., and Werner, M. (1995), The SIR-C/X-SAR synthetic aperture radar system, *IEEE Trans. Geosci. and Remote Sensing* 33: 829-839.

Lang, R.H., and Sidhu, J.S. (1983), Electromagnetic backscattering from a layer of vegetation: a discrete approach, *IEEE Trans. Geosci. Remote Sensing* 21: 62-71.

Meneghini, R., and Atlas, D.A. (1986), Simultaneous ocean cross section and rainfall measurements from space with a nadir-looking radar, *J. Atmos. Ocean. Tech.* 3: 400-413.

Newell, R.E., and Geotis, S.G. (1955), Meteorological measurements with a radar provided with variable polarization, *Weather Research Tech.* Note No. 8, Dept. of Meteorology, Massachusetts Institute of Technology, 12 pp.

Okamoto, K., Awaka, J., and Kozu, T. (1988), A feasibility study of rain radar for the tropical rainfall measuring mission: 6. A case study of rain radar system, *J. Comm. Research Lab.* 35: 183-208.

Olsen, R.L., Rodgers, D.V., and Hodge, D.B. (1978), The aR^b relation in the calculation of rain attenuation, *IEEE Trans. Ant. Prop.* 26: 318-329.

Phillips, O.M. (1977), *The Dynamics of the Upper Ocean*, Cambridge University Press, Cambridge, UK.

Poon, Y.K., Tang, S., and Wu, J. (1992), Interactions between rain and wind-waves, *J. Phys. Ocean.* 22: 976-987.

Simpson, J., Adler, R.F., and North, G. (1988), A proposed tropical rainfall measuring mission (TRMM) satellite, *Bull. Amer. Meteor. Soc.* 69: 278-295.

Stofan, E.R., Evans, D.L., Schmullius, C., Holt, B., Plaut, J.J., van Zyl, J.J., Wall, S.D., and Way, J. (1995), Overview of results of Spaceborne Imaging Radar-C, X-band Synthetic Aperture Radar (SIR-C/X-SAR), *IEEE Trans. Geosci. and Remote Sensing* 33: 817-828.

Tsimplis, M.N. (1992), The effect of rain in calming the sea, *J. Phys. Ocean.* 22: 404-412.

Ulaby, F.T., Held, D., Dobson, M.C., McDonald, K.C., and Senior, T.B.A. (1987), Relating polarization phase difference of SAR signals to scene properties, *IEEE Trans. Geosci. Remote Sensing* 25: 83-92.

Ulbrich, C.W. (1983), Natural variations in the analytical form of the raindrop size distribution, *J. Climate Appl. Meteor.* 22: 1764-1775.

Figure captions

Figure 1: GMS IR images of Typhoon Seth with the Shuttle tracks overlaid: (a) October 6, 1994, 17:32 GMT – the two crosses mark the location of the convective cells in figures 2 and 3; (b) October 3, 1994, 17:32 GMT – the cross marks the location of the stratus in figure 4.

Figure 2: Along track cross-section of the radar reflectivities from a convective cell in Typhoon Seth. Vertical tick marks indicate 1-km intervals, horizontal ones 3-km intervals. The surface is represented in navy blue. Note the two different color scales for the X/C and L-bands.

Figure 3: Along track cross-section of the radar reflectivities from a second convective cell in Typhoon Seth. Vertical tick marks indicate 1-km intervals, horizontal ones 3-km intervals. The surface is represented in navy blue. The color scales for the C-VV and L-bands are as in figure 2. C-VH is represented using the third color scale shown at the bottom.

Figure 4: Along track cross-section of the radar reflectivities from stratiform precipitation in Typhoon Seth. Vertical tick marks indicate 1-km intervals, horizontal ones 3-km intervals. The surface is represented in navy blue. The three different color scales for X/C-VV, L, and C-VH are as in figures 2 and 3.

Figure 5: Profiles of radar parameters through melting layers as a function of height above the surface on three different occasions using three different radars: (a) Spaceborne SIR-C/X-SAR measurements of the linear depolarization ratio (LDR) at C-band and the radar reflectivity factors (Z) at L- C- and X-bands associated with Typhoon Seth; (b) Airborne ARMAR measurements including the profile of particle vertical velocities (v=fallspeed + vertical air speed) in a developing tropical storm over the Western Pacific; (c) Measurements in Florida using the ground- based NCAR CP-2 radar.

Figure 6: Reflectivity profiles through the convective cell in figure 2.

Figure 7: Retrieved precipitation profiles: (a) rainrate with 1-σ uncertainty (liquid water can extend to significant heights, but it is impossible to know how far above 5-6 km the profile should be trusted); (b) DSD parameter Λ.

Figure 8: Three-frequency VV images from DT 13.4 during SRL-1, April 10, 1994, 05:36 GMT: (a) X-band; (b) C-band; (c) L-band. The bright linear feature to the right of the images is part of the Ontong Java atoll in the Solomon Islands.

Figure 9: Contrast-enhanced version of the L-band panel in figure 9 with dark areas noted A-G.

Figure 10: Three-frequency VV images of rain cells over the Amazon from DT 103.6 during SRL-1: (a) X-band; (b) C-band; (c) L-band.

Figure 11: Scattering mechanisms from a random medium over a reflecting surface.

Figure 12: Observed and modeled dependence of σ^o versus range for an Amazon rain cell in DT 103.6 .