

MICROWAVE RADIOMETER STUDIES OF ATMOSPHERIC WATER OVER THE OCEANS

VOL. II:

Technical Report for NASA Grant NAGW-1688

- * Petty, G.W. and K.B. Katsaros, 1990: Precipitation observed over the South China Sea by the Nimbus 7 Scanning Multichannel Microwave Radiometer during Winter MONEX. *J. Appl. Meteor.*, **29**, 273-287.
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MICROWAVE RADIOMETER STUDIES OF ATMOSPHERIC WATER OVER THE OCEANS

A Technical Report in Two Volumes

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INTRODUCTION

Since the Seasat satellite carried the Scanning Multichannel Microwave Radiometer (SMMR) into space in July of 1978, shortly followed by the SMMR on Nimbus 7, which operated for almost a decade, a new type of data source on atmospheric water vapor and other meteorological parameters has been available for analysis of weather systems over the ocean. Since 1987 we have had the Scanning Multichannel Microwave/Imager (SSM/I) instrument on Defense Meteorological Satellites providing similar data. In this two volume technical report we present a collection of our work performed over the last three years under NASA grant NAGW-1688 administered by Dr. James Dodge of the Mesoscale Program at NASA Headquarters, and NASA grant NAG5-943 administered by Dr. Ramesh Kakar of the Global Processes Program. Some of the chapters are journal articles, others are conference proceedings and abstracts. A major contribution from our group is the Ph.D. thesis of Grant W. Petty, 1990, produced separately as Final Technical Report on NASA grant NAG5-943. It is available upon request.

Volume I of this report contains work mainly supported by NASA grant NAGW-1688. The first is an article for publication in *Monthly Weather Review*:

Miller, D.K. and K.B. Katsaros, 1992: Satellite derived surface latent heat fluxes in rapidly intensifying marine cyclones. *Mon. Wea. Rev.*, (scheduled publication: June).

It is derived from the following Masters of Science thesis:

Miller, Douglas K., 1990: *Estimation of Surface Latent Heat Flux Patterns Preceding a Rapidly Intensifying Cyclone Derived from the Special Sensor Microwave/Imager*. M. Sci. Thesis, Department of Atmospheric Sciences, University of Washington, Seattle, WA 98195. [Not included in this report.]

In this thesis a method for evaluating the preconditioning of the marine environment via surface moisture flux 24 hours in advance of rapid deepening of a cyclonic storm is explored with data from a 1988 pre-ERICA (Extratropical Rapidly Intensifying Cyclones over the Atlantic) storm. We plan to carry on this project with studies of the 1989-90 ERICA storms (Hadlock and Kreitzberg, 1988).

One of our goals was to make full use of the new tool for studying marine weather systems that the SSM/I represents by applying our algorithms and versatile graphics presentation to the task at hand. The hardware and software system were developed by G.W. Petty while a graduate student with funds supplied by a special NASA grant (NAS8-36473, Cyclonic Storms).

Collaboration with the following colleagues and visiting student researchers have enabled us to advance along several avenues at once: Dr. Nelly Mognard at the University of Puget Sound; Dr. Chantal Claud of Ecole Polytechnique Paliseau, student of Professor Alain Chedin; and Ms. Aurelie Sand of the University of Paris VII, student of Professor Claude Klapisz. Results of the work presented at the European Geophysical Society meeting in Copenhagen, April 23-27, 1990, were reported on in our Third Annual Report, 1990. The following new work was presented at the Royal Meteorological Society/American Meteorological Society joint meeting in London, September 3-7, 1990:

Claud, C., N. Scott, A. Chedin, K. Katsaros and G. Petty, 1990: Mesoscale Meteorology in Polar Regions from a Coupling of NOAA (TOVS) and DMSP (SSM/I) Data. In Preprint Volume, *Fifth Conference on Satellite Meteorology and Oceanography*, September 3-7, 1990, London, U.K. Published by the Am. Meteorol. Soc., Boston, MA.

Mognard, N.M. and K.B. Katsaros, 1990: Evolution of Atmospheric Fronts over the Ocean as Observed with the Special Sensor Microwave/Imager and the Geosat Altimeter. In Preprint Volume, *Fifth Conference of the Royal Meteorological Society and the American Meteorological Society*, London, U.K., September 3-7, 1990.

The following article which will appear in *Tellus* is an extension of the London presentation by Claud et al. (1990):

Claud, C., K.B. Katsaros, G.W. Petty, A. Chedin and N.A. Scott, 1992: A cold air outbreak over the Norwegian Sea observed with the Tiros-N Operational Vertical Sounder (TOVS) and the Special Sensor Microwave/Imager (SSM/I). *Tellus*, (in press).

Follow-up papers to the conference presentations by Mognard, and by Sand, are also being prepared.

Earlier work supported by the Mesoscale Program of NASA appeared this year as a journal article:

McMurdie, L.A. and K.B. Katsaros, 1991: Satellite-derived precipitable water distribution in oceanic midlatitude storms: variation with region and season. *Mon. Wea. Rev.*, **119**, 589-605.

It is also included in this report, Vol. I. Collaborative work with Scandinavian colleagues in relating the SMMR or SSM/I water parameters to similar quantities extracted from numerical models of the atmosphere have resulted in a University of Stockholm Technical Report:

Katsaros, K.B., G.W. Petty and U. Hammarstrand, 1990: Atmospheric Water Parameters in Mid-Latitude Cyclones Observed by Microwave Radiometry and Compared to Model Calculations. *Technical Report*, Meteorological Institute, University of Stockholm (in press). [Not included in this collection but available on request.]

An extension of ideas for this report were carried out with a later version of the Stockholm-Bergen mesoscale atmospheric model. A short article for *Tellus* is included in this report, Vol. I:

Raustein, E., H. Sundqvist and K.B. Katsaros, 1991: Quantitative comparison between simulated cloudiness and clouds objectively derived from satellite data. *Tellus*, **43A**, 306-320.

Volume II of this technical report contains work mainly supported by NASA grant NAG5-943 and its predecessor NASA grant NAG5-354. We recognized early on the importance of microwave radiometer estimates of precipitation over the oceans, but also the difficulty in evaluating the accuracy of such estimates since few comparison data exist. Radar data

provide a source for comparison, but the difference in resolution between the satellite estimates of the order of 35 km and the radar estimates of the order of 0.5 km are a problem. Two articles concerned with these problems have been produced and are reproduced here:

Petty, G.W. and K.B. Katsaros, 1990: Precipitation observed over the South China Sea by the Nimbus 7 Scanning Multichannel Microwave Radiometer during Winter MONEX. *J. Appl. Meteor.*, **29**, 273-287.

Petty, G.W. and K.B. Katsaros, 1992: Nimbus 7 SMMR precipitation observations calibrated against surface radar during TAMEX. *J. Appl. Met.*, (scheduled publication: May).

The research performed by Petty was also presented at the American Meteorological Society Satellite Conferences in San Diego, 1989, and London, 1990. The London conference proceedings are:

Petty, G.W. and K.B. Katsaros, 1990: New geophysical algorithms for the Special Sensor Microwave/Imager. Preprint Volume, *Fifth International Conference on Satellite Meteorology and Oceanography*, September 3-7, 1990, London, England. [Included in this report.]

Petty, G.W. and K.B. Katsaros, 1990: Nimbus 7 SMMR precipitation observations calibrated against surface radar during TAMEX. *J. Applied Met.*, (submitted). Also appearing in abbreviated form in Preprint Volume, *Fifth International Conference on Satellite Meteorology and Oceanography*, September 3-7, 1990, London, England. [Not included here; overlaps with the article above with the same title.]

The thesis work of Petty will appear as a series of articles in the near future. One has been submitted.

A review of the legacy that the Seasat Satellite provided for all of the work in our group on microwave radiometry and other work related to the scatterometer and altimeter instruments has appeared in the *Bulletin of the American Meteorological Society*:

Katsaros, K.B. and R.A. Brown, 1991: Legacy of the Seasat Mission for studies of the atmosphere and air-sea-ice interactions. *Bull. Am. Met. Soc.*, **72**, 967-981.

Three papers are accepted for presentation at the Sixth International Conference on Satellite Meteorology and Oceanography, in early 1992.

McMurdie, L. and K.Katsaros, 1992: Satellite Derived Integrated Water Vapor and Rain Intensity Patterns: Indicators of Rapid Cyclogenesis. In Preprint Volume, *6th AMS Conference on Satellite Meteorology and Oceanography*, January 5-10, 1992, Atlanta, GA.

Mognard, N.M. and K.B. Katsaros, 1992: Comparison of Wind Speed Measurements over the Oceans with the Special Sensor Microwave/Imager and the Geosat Altimeter. In Preprint Volume, *6th AMS Conference on Satellite Meteorology and Oceanography*, January 5-10, 1992, Atlanta, GA.

Petty, G.W. and K.B. Katsaros, 1992: Morning-Evening Differences in Global and Regional Oceanic Precipitation as Observed by the SSM/I. In Preprint Volume, *6th AMS Conference on Satellite Meteorology and Oceanography*, January 5-10, 1992, Atlanta, GA.

The preprints of these conference papers are the final contribution in Volume II.

References:

Abbey, R.F., C.W. Kreitzberg and R. Hadlock, 1987: Experiment on Rapidly Intensifying Cyclones over the Atlantic, ERICA, Overview Document, Field Research Project, Version 1. March 1987, available from the ERICA Project Office, Battelle Memorial Institute, Richland, WA.

Hadlock, R. and C.W. Kreitzberg, 1988: The Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA) field study-objectives and plans. *Bull. Am. Met. Soc.*, **69**, 1309-1320.

Petty, G.W., 1990: On the Response of the Special Sensor Microwave/Imager to the Marine Environment - Implications for Atmospheric Parameter Retrievals. Ph.D. Thesis, October 1990, Dept. of Atmospheric Sciences, University of Washington, Seattle, WA 98195. NASA Tech Report.

Precipitation Observed over the South China Sea by the Nimbus-7 Scanning Multichannel Microwave Radiometer during Winter MONEX

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(Manuscript received 8 August 1988, in final form 28 May 1989)

ABSTRACT

Mesoscale cloud clusters near the northwestern coast of Borneo were observed by the Scanning Multichannel Microwave Radiometer (SMMR) on three occasions during the Winter Monsoon Experiment in December 1978. A nondimensional form of the SMMR 37 GHz polarization difference is introduced and used to identify regions of precipitation, and these are compared with visible and infrared imagery from the GMS-1 geostationary satellite. For two of the three cloud cluster cases, quantitative comparisons are made between nearly simultaneous SMMR observations and reflectivity observations made by the MIT WR-73 digital weather radar at Bintulu. Though limited in scope, these represent the first known direct comparisons between digital radar-derived rain parameters and satellite passive microwave observations of near-equatorial precipitation. SMMR 37 GHz observations are found to be much better indicators of fractional coverage of each SMMR footprint by rain than of average rain rate within the footprint. Total area coverage by precipitation is estimated for all three clusters using this result.

1. Introduction

Precipitation over the oceans is an important meteorological and oceanographic quantity which is very difficult to measure directly. In the tropics it is of particular interest, because it is a measure of the strong warming of the atmosphere at these latitudes by the heat of condensation. This heating is an important term in atmospheric circulation models (e.g., Ramage 1968; Manabe et al. 1970; Webster 1972; Hartmann et al. 1984).

The bulk of all tropical precipitation occurs in tropical cloud clusters, which are evident in satellite imagery by their mesoscale cirriform cloud shields. Within these cloud clusters, most of the rain falls in mesoscale precipitation features (MPFs) composed of a combination of convective and stratiform components. Although tropical cloud clusters are initiated and driven by convective activity, there is evidence that the stratiform component may actually contribute between a third and half of the total accumulated rainfall in tropical cloud clusters (Houze and Betts 1981; Houze and Hobbs 1982). Satellite techniques which exploit only indirect indices of precipitation—often based on purely convective models of tropical rainfall—may have difficulty providing reliable estimates of this contribution, since the visible and infrared signature of stratiform rain clouds can be very similar to that of nonraining clouds.

Passive microwave techniques for sensing precipitation have been in experimental use for over a decade (e.g., Rao and Theon 1977; Wilheit et al. 1977; Weinman and Guetter 1977; Jung 1980; Spencer et al. 1983; McMurdie and Katsaros 1985; Katsaros and Lewis 1986; Spencer 1986). The principal appeal of microwave techniques, as contrasted with those which sense only cloud top parameters, is that the microwave brightness temperature of a rain cloud changes in direct response to the *vertically integrated* liquid water constituents of the cloud, especially precipitation-size drops.

The primary difficulty with satellite-borne microwave sensors to data, particularly with regard to the sensing of rain, has been the relatively poor surface resolution achievable using practical antenna sizes. For a given antenna diameter, however, spatial resolution improves with increasing radiometer frequency. Thus, for obtaining spatial distributions of precipitation, channels of 18 GHz or greater have generally been favored. Higher frequencies are also more sensitive to light rain, although a penalty is paid in the form of smaller dynamic range and greater nonlinearity of the response.

The Scanning Multichannel Microwave Radiometer (SMMR) flown on Seasat and Nimbus 7 (Gloersen and Barath 1977) is equipped with ten channels consisting of five frequencies ranging from 6.6 GHz (150 km resolution) to 37 GHz (30 km resolution) in horizontal and vertical polarization. The SMMR was intended for the retrieval of integrated water vapor, cloud liquid water, sea surface temperature, ice coverage,

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surface wind speed and rain rate. In the following, we restrict our attention to the two 37 GHz channels for the estimation of rain parameters.

Studies of the atmospheric monsoon circulations over the Asian continent were carried out in conjunction with the First GARP (Global Atmospheric Research Programme) Global Experiment (FGGE). One regional experiment within FGGE was the Winter Monsoon Experiment (WMONEX), carried out over the South China Sea in December 1978 (Greenfield and Krishnamurti 1979). Though limited in quantity and scope, the digital radar observations of MPFs obtained in this experiment provide some of the first independent comparison data for SMMR observations of near-equatorial precipitation.

In this study, we compare SMMR observations of three tropical cloud clusters during WMONEX with operational visible and infrared satellite imagery and with radar estimates of precipitation parameters. Our principal aim is to help demonstrate the usefulness of passive microwave data for the global monitoring of rainfall and to begin building a semiempirical basis for the quantitative interpretation of microwave observations of tropical oceanic cloud clusters. In support of this aim, we introduce a normalized form of the observed 37 GHz brightness temperature polarization difference which we interpret in terms of microwave attenuation by rain, as contrasted with previously published emission-based (e.g., Wilheit et al. 1977) and scattering-based algorithms (Spencer 1986).

In the next section, we briefly review the physical basis for monitoring rainfall with satellite passive microwave data, with particular attention to the motivation for an attenuation-based approach. Section 3 describes the processing of the digital radar and the SMMR data which were employed in the analysis; in section 4, the three cloud cluster cases over the South China Sea are discussed. In section 5, digital radar reflectivities obtained for two of the cases are compared with nearly simultaneous SMMR observations of the same region.

2. Rain parameter retrieval from 37 GHz brightness temperatures

a. Theory

The ability of the SMMR to detect rainfall over the ocean is due to the strong contrast between the thermal microwave emission of rain and that of the sea surface (e.g., Wilheit et al. 1977; an early prediction of this possibility was made by Buettner 1963). At 37 GHz, the microwave emissivity (ϵ) of the ocean is low (~ 0.5) so that the ocean provides a relatively "cold" radiometric background when seen from space, whereas the bulk emissivity of an optically thick layer of rain is closer to unity. Rain over the ocean therefore produces dramatically higher satellite-observed brightness temperatures.

In addition to emission effects, there is also some scattering of microwave radiation by raindrops and precipitation-size ice particles. This scattering effect increases with increasing particle size, as shown by Mie theory calculations (e.g., Diermendjian 1969; Savage and Weinman 1975), and acts to depress brightness temperatures below those which would occur in a layer of equal opacity but without scattering. Hence, as rain rate increases, the brightness temperature first increases rapidly and then "saturates" for some critical value of rain rate, typically near 4 mm h^{-1} for a 4 km deep layer. This is followed by a reduction in brightness temperature with further increases in rain rate, as more large rain drops and frozen precipitation particles aloft influence the radiative properties of the cloud (Savage 1976; Wu and Weinman 1984; Spencer 1986).

Other geophysical factors influence 37 GHz brightness temperatures, albeit to a lesser degree than rain. The most important of these are atmospheric water vapor content, cloud liquid water, sea surface temperature and wind-induced roughness (Wilheit and Chang 1980; Wentz 1983). In order for a rain rate retrieval algorithm to give reliable results under the most general conditions, particularly for lower rain rates, the variability of these parameters should be taken into account.

Most passive microwave precipitation retrieval techniques to date (e.g., Wilheit et al. 1977) have attempted to exploit the emissive characteristics of rain clouds by relating changes in brightness temperature at one polarization to variations in rain rate. More recently, Spencer (1986) has proposed a semiempirical technique which relates the observed scattering properties of deep convective clouds at 37 GHz to rain intensity over the ocean. It is primarily intended for the retrieval of large rain rates associated with strong convective clouds containing significant numbers of large frozen precipitation particles (i.e., graupel, hail) above the freezing level, and may thus prove to be a useful complement to emission-based algorithms for light rain. The method, however, was derived primarily from observations of intense storms over the North American continent, where the relationship between concentrations of ice particles aloft and surface rain rate may not be representative of convection elsewhere. For example, LeMone and Zipser (1980) and Zipser and LeMone (1980) found considerably lower maximum updraft velocities in tropical oceanic convection than are typical in thunderstorms over North America. This result suggests that there may also be proportionally fewer of the largest ice particles to which the scattering-based method is most sensitive. Investigations employing direct observations of tropical convective clouds are therefore still necessary.

In this study, we pursue a third approach to interpreting passive microwave observations of rain. This method exploits the effect of attenuation by rain on the polarization of the observed microwave brightness

temperatures. At the oblique viewing angle (50.3°) of the SMMR on Nimbus 7, the microwave sea surface emissivity is greater for vertical than for horizontal polarization, so that, over the unobscured ocean, there is typically a 40–70 K difference between the measured 37 GHz brightness temperatures in the two polarizations. But the atmosphere, including rain, both attenuates and emits nearly independently of polarization, so that the brightness temperature polarization difference (defined here as $T_{B,V} - T_{B,H}$, where the subscripts V and H refer to vertical and horizontal polarization) associated with rain over the ocean is much smaller. The difference decreases monotonically to near zero with decreasing visibility of the ocean surface (Beer 1980; Spencer et al. 1983). In the following, we offer a refinement to this interpretation, by expressing the polarization difference as a fraction of the value expected in the absence of rain and cloudiness. The result can be interpreted in a relatively straightforward way in terms of an effective rain cloud transmittance, with much of the influence of unrelated geophysical parameters having been factored out.

The main limitations inherent in any purely theoretical method for rain rate estimation stem from the coarse spatial resolution of the SMMR and the strong spatial inhomogeneity and complex microphysical structure typical of rain clouds. It has therefore proven difficult to obtain unbiased rain rate retrievals simply by inverting relationships derived from simple models of the atmosphere. Until a great deal more is known about rain cloud structure and statistics on a global basis, significant improvements in theoretical relationships are unlikely. Thus, although a simple model will suffice to qualitatively explain the polarization signature of rain, we must rely on data from surface based radars to develop useful empirical relationships between SMMR parameters and tropical precipitation.

b. Depolarization of microwave brightness temperatures by rain

We define a new parameter, the *normalized polarization difference*, P , as the difference between the vertically and horizontally polarized brightness temperature ($T_{B,V} - T_{B,H}$), divided by a hypothetical *clear sky* polarization difference ($T_{B,V} - T_{B,H}$)_{CLR}:

$$P = \frac{(T_{B,V} - T_{B,H})}{(T_{B,V} - T_{B,H})_{\text{CLR}}} \quad (1)$$

By definition, P equals unity when no clouds are present. When sufficiently heavy rain is present throughout the satellite field of view (FOV), P approaches zero. Thus, P can be thought of as a measure of the visibility of the sea surface through or between clouds and rain as seen by the microwave sensor in space, relative to a cloudless atmosphere under otherwise similar conditions.

Theoretically, heavy precipitation could be associated with nonzero P when large, preferentially oriented, aspherical scatterers (e.g., rimed ice plates or dendrites) are present in sufficient concentrations (Wu and Weinman 1984). Although this effect might introduce ambiguities into the interpretation of P under certain circumstances, we have found no evidence that this is important in the oceanic tropical cloud clusters treated here.

For the ideal situation in which rain and cloud in the FOV are uniformly distributed in a horizontal layer, modeling studies by the authors indicate that P , under rather general conditions, is approximately equal to the square of the slant-path transmittance τ through the raincloud at the viewing angle θ of the SMMR. Thus,

$$P \approx \tau^2 = \left\{ \exp \left[\frac{-\sigma}{\cos \theta} \right] \right\}^2 = \exp \left[\frac{-2\sigma}{\cos \theta} \right], \quad (2)$$

where σ is equal to the sum of the optical depth of the rain (σ_R) and that of nonprecipitating cloud water (σ_L). The σ_R can in turn be approximated as the product of an effective depth of the rain layer and the volume extinction coefficient, $k_R(R)$ (km^{-1}), where R is the rain rate. At 37 GHz, k_R varies nearly linearly with rain rate (Savage 1976), and σ_L is proportional to the vertically integrated cloud water mass per unit area (comprised of droplets of radius less than $50 \mu\text{m}$).

If one knew both the effective rain layer depth and the relative contributions of cloud liquid water and rain to the rain cloud total opacity, one could obtain a reasonable quantitative estimate of the rain rate directly from observations of P using (2), provided both that the condition of horizontal homogeneity holds and that the optical depth σ does not greatly exceed unity. At 37 GHz, a 4 km thick rain layer attains unit optical depth for rain rates of approximately $3\text{--}4 \text{ mm h}^{-1}$.

The assumption of horizontal homogeneity, however, is seldom valid on the 30 km scale of the SMMR 37 GHz FOV. Thus, the SMMR-observed value of the normalized polarization difference can be more realistically interpreted as an average over the FOV of the polarization difference at each point. If the latter is taken to be related to the local rain rate by (2), then the rain rate retrieved from observed (i.e., spatially averaged) values of P by directly inverting (2) will, in general, be an underestimate of the true value of $\bar{R}$, the average rain rate within the satellite FOV. This is simply because the sensitivity of P to changes in rain rate decreases with increasing rain rate; i.e., the first and second derivatives of P with respect to R have opposite signs. For fixed $\bar{R}$, the underestimate is increasingly serious with increasing inhomogeneity of the rain in the FOV; e.g., as more of the rain is concentrated in small, intense convective cores or narrow rainbands. In the case of extreme inhomogeneity, i.e., when most of the area of the FOV consists of regions of no rain

together with regions of optically thick rain it can be readily shown that the observed value of P for a SMMR pixel may be simply related to the fractional coverage F of the FOV by rain:

$$P \approx (1 - F). \quad (3)$$

Although the present discussion has focused on the normalized polarization difference, P , analogous problems arise in the interpretation of the raw brightness temperatures. No matter which method is used, there is always some unavoidable ambiguity in the purely physical interpretation of SMMR 37 GHz observations when a priori knowledge of the spatial distribution of rain within the FOV is unavailable, as is almost always the case with the present satellite instruments. In order to minimize statistical biases in the estimates due to such uncertainties, high priority should be given to developing and exploiting empirical relationships between satellite observables and the meteorological parameters of interest.

Our motivation for using P as an independent variable in such relationships rests on its following advantages: first, it allows simple (yet distinct) quantitative physical interpretations in the horizontally homogeneous case and in the case of extreme inhomogeneity. Second, it does not share the nonmonotonic response of brightness temperature to rain rate. Most importantly, variations in P are expected to be more directly related to rain effects alone. This is so because changes in microwave signature due to large-scale variability of sea surface emissivity, sea surface temperature, and atmospheric opacity due to gaseous components are, to a great degree, factored out when normalizing by the clear-sky polarization difference. The latter, we have assumed, can be adequately estimated within concentrated regions of rain using microwave observations of the surrounding area. One technique for doing this is described in section 3b.

3. Data processing

a. Radar

During WMONEX, which was conducted in December 1978 in the South China Sea, the Massachusetts Institute of Technology WR-73 C-band digital weather radar was in operation at Bintulu on the north coast of Borneo (Houze et al. 1981a,b). Characteristics of the radar are listed in Table 1. The radar operated during the entire period without significant interruption, sampling a volume of the atmosphere at a sequence of elevation angles out to a nominal range of 256 km from Bintulu. Each volume scan required approximately 8 minutes to complete, and a new sequence was initiated every 10 minutes. Unfortunately, the site was affected by partial blockage of the radar beam by trees and buildings at low elevation angles, and particularly for some azimuths. Therefore, some care is nec-

TABLE 1. Characteristics of MIT WR-73 digital radar located at Bintulu during WMONEX.

Wavelength	5.3 cm
Pulse length	300 m
PRF	250 s ⁻¹
Peak power	250 kW
Beam width	1.45°
Azimuth averaging interval	1.0°
Azimuth recording interval	1.0°
Minimum detectable signal	-105 dBm
Range bin size	0.25-1.0 km
Maximum range recorded	256 km

essary in interpreting the recorded reflectivity patterns (see section 5a).

For this study, the volume scans of interest were interpolated to Cartesian coordinates using a program which employed a Cressman interpolation scheme and corrected for the earth's curvature and refraction under standard atmospheric conditions. Cartesian grid spacing was chosen to be 3.75 km, which corresponds to one-eighth of the diameter of a gridded 37 GHz SMMR pixel. The vertical grid spacing was one kilometer. Because of the problem of low-elevation beam blockage, reflectivities in at least the lowest kilometer were unreliable, particularly near the limits of the radar range. To obtain reasonable estimates of rain rate at each horizontal grid point, the largest reflectivity value above the surface but below 4 km was taken to be representative. The latter constraint was employed to avoid contamination by bright band effects near the freezing level, which, in these cases, typically occurred near a height of 4.5 km above sea level.

To convert radar reflectivity values to rain rate, we employed the Z - R relationship of Austin and Geotis (1979), which was derived from observations of near-equatorial oceanic precipitation during the GARP (Global Atmospheric Research Programme) Atlantic Tropical Experiment (GATE):

$$Z = 180 R^{1.35}. \quad (4)$$

b. SMMR

The Nimbus 7 SMMR data consist of brightness temperatures at two polarizations (vertical and horizontal) for each of the five SMMR frequencies, remapped onto a rectangular grid of uniformly spaced cells whose size depended on the frequency. At 37 GHz, the spacing is approximately 30 km both along and across the swath, and the total swath width is 26 cells or approximately 780 km. The slight loss of resolution resulting from the remapping process is offset by the added convenience of the uniform spacing.

The navigation of the SMMR data was checked by comparing the 37 GHz horizontally polarized brightness temperatures against the northern and southern Borneo coastlines. North-south navigation corrections of up to 0.7 cell diameters, or about 20 km, were nec-

essary for some orbits in order to bring the SMMR images of the Borneo coastline into agreement with its true position.

The baseline (clear-sky) polarization difference ($T_{B,V} - T_{B,H}$)_{CLR} in (1) was determined for each SMMR pixel from a square array of pixels (dimension $M \times M$, where M is an odd integer) centered on the pixel in question. After rejecting all pixels within the block which showed a polarization difference of less than 35 K, indicating unambiguously that they were influenced by clouds, rain or land, the N th percentile value of the remaining polarization differences was chosen as the baseline polarization difference.

The choice of M and N is subjective but should reflect the horizontal scale of the rain clouds being examined as well as the fraction of SMMR pixels within the $M \times M$ block which are expected to correspond to clear or nearly clear sky conditions. If M is too small, a pixel block may not contain enough cloud-free pixels; if it is too large, the clear-sky pixels selected may not represent ambient conditions at the center of the block.

After some experimentation, we found that $M = 13$ and $N = 90$ led to smooth and physically consistent fields of the clear-sky polarization difference for our cases. Values typically ranged from approximately 42 K near the Borneo coast to 55 K near the coast of the People's Republic of China. The difference is primarily due to a north-south gradient in the total atmospheric water vapor content. Indeed, the clear-sky polarization field agrees, to within two or three degrees, with the values predicted by the model of Wentz (1983) applied to the water vapor and wind speed fields obtained using the lower frequency SMMR channels.

The observed polarization differences for each pixel were divided by the corresponding clear-sky value to yield the field of P . Comparison of maps of P to GMS-1 imagery of the area revealed that nearly cloud-free areas were associated with values of P close to unity, as expected. Occasionally, pixels with $P > 1$ appear, especially near the Borneo coast. For these pixels, the technique described above apparently underestimates the clear-sky polarization difference. Using a larger value for N would reduce the number of pixels for which the clear-sky polarization is underestimated, but might also lead to the entire clear-sky polarization field being determined by a very small number of pixels exhibiting anomalously high polarization. Such pixels might occur, for example, in clear-sky regions where wind roughening of the surface is locally reduced due to the influence of the coast. In any case, small errors—say, 2 or 3 degrees—in the clear-sky polarization introduce errors in P which are small compared with the variations due to precipitation.

The applicability of this particular method for estimating the clear-sky polarization difference depends both on the fields of water vapor, surface roughness, and sea surface temperature being relatively uniform on a horizontal scale comparable to that of the cloud

cluster, and on the region of precipitation itself being small in comparison to the 780 km swath width of the SMMR. In other situations, the clear-sky polarization difference might be better estimated using independent meteorological or climatological information combined with a suitable brightness temperature model (e.g., Wentz 1983).

Figure 1 depicts a plot of P vs horizontally polarized brightness temperature $T_{B,H}$ for the three cases discussed below, covering the oceanic region north of Borneo and south of 10°N . The $T_{B,H}$ is the more familiar microwave index of rain; it is also highly correlated with P —high brightness temperatures due to emission by cloud are associated with low polarizations due to attenuation of the sea surface signal. The scatter in the plot can be attributed to (i) variations in the background which are factored out in P but not in $T_{B,H}$ and (ii) differences in the effective emitting temperature of the rain clouds in each pixel, including, in some cases, those due to volume scattering by ice.

4. Cases

Unlike the short-lived, propagating convective disturbances observed during GATE (Leary and Houze 1979) and routinely in many other regions of the world as well, mesoscale cloud clusters studied in WMONEX were slow moving and could persist in a given area for periods of a full day or longer. They were generally large, often possessing horizontal scales of several hundred kilometers. Houze et al. (1981b) found that convection in WMONEX was subject to a well-defined diurnal cycle, with convective activity offshore peaking in the early morning hours, due to interaction of the large scale monsoon flow with local land-breeze circulations. By 0800 LST, convection had typically merged into an organized mesoscale system, with numerous convective cores bound together by an extensive (~ 200 km) stratiform precipitation area. By mid-day, the mesoscale systems usually began dissipating, so that stratiform rainfall from the large mid-to-upper level cloud shield increasingly replaced convection as the dominant source of precipitation.

For a more detailed discussion of the climatology and meteorology of tropical cloud clusters over the South China Sea during WMONEX, the reader is referred to Houze et al. (1981b), Johnson and Priegnitz (1981), Churchill and Houze (1984), and other references cited therein.

We examined cloud clusters in the South China Sea on 11, 17, and 23 December 1978, the first and the last dates being of particular interest because of nearly simultaneous, overlapping coverage by the SMMR and the MIT radar. The coincident coverage enabled us to make a preliminary assessment of P as an index of tropical precipitation activity. The cloud cluster on 17 December was not detected by the radar at the time of the SMMR overpass, despite lying partially within the radar's nominal range of 256 km.

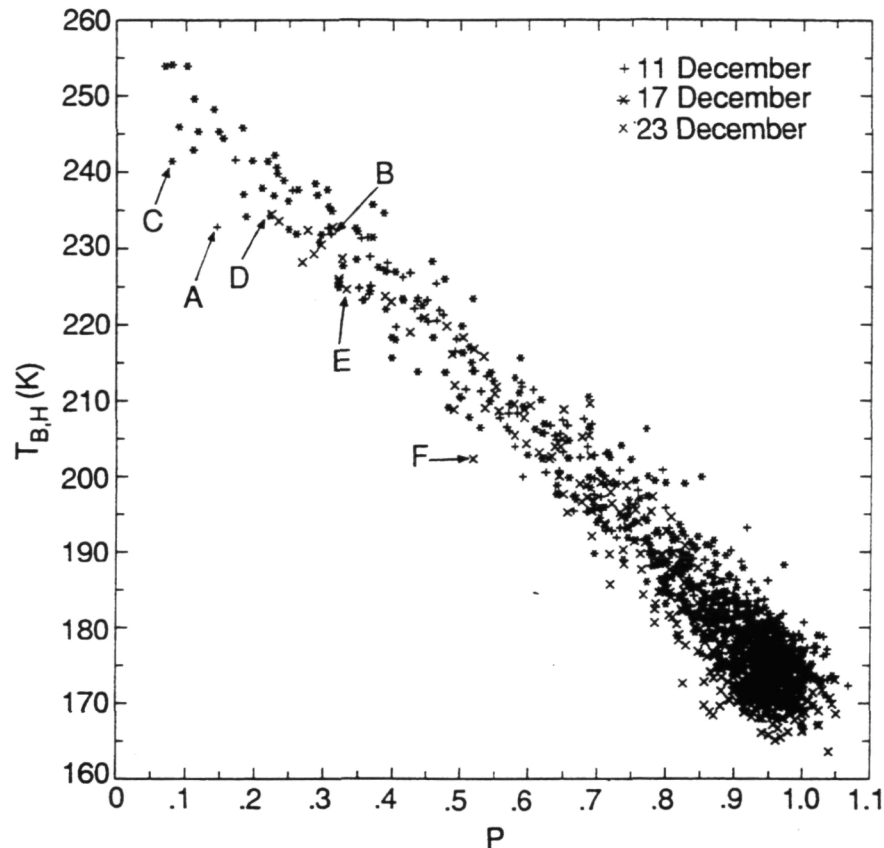


FIG. 1. Scatter plot of SMMR normalized polarization difference (P) vs horizontally polarized brightness temperature ($T_{B,H}$) at 37 GHz, for the three cloud cluster cases discussed in the text. Data points labeled with letters correspond to lettered SMMR pixels in Figs. 2–4.

The 11 and 23 December cloud clusters were observed on days of unusually strong northeasterly monsoonal flow over the South China Sea associated with so-called “cold surges,” which frequently bring heavy rainfall to the North Borneo coast. The 17 December cluster occurred during the period of declining northeasterly flow following a strong cold surge on 15 December.

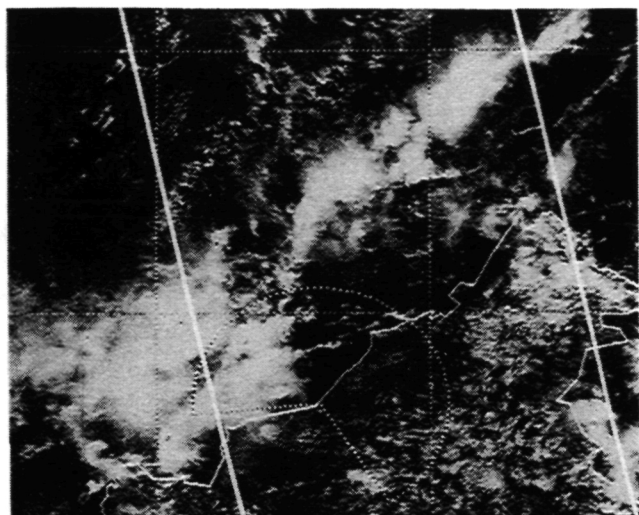
In each case discussed below, the cluster was observed during the daytime leg of the Nimbus 7 orbit, which always occurred near noon LST (0400 UTC on Borneo). The observations thus generally coincide with the weakening phase of the clusters, although new convection is still observed in two of the cases.

Visible and infrared (IR) images from a Japanese geostationary meteorological satellite, GMS-1, were obtained for 0233 and 0533 UTC, roughly 1.5 hours before and after each SMMR overpass. Because of the time differences, only qualitative comparisons of the GMS-1 and the SMMR images are possible. In order to facilitate such comparisons and to best convey the evolution of the cloud clusters, the GMS-1 visible and IR images are remapped and displayed in the same geographic projection as the SMMR P fields in Figs.

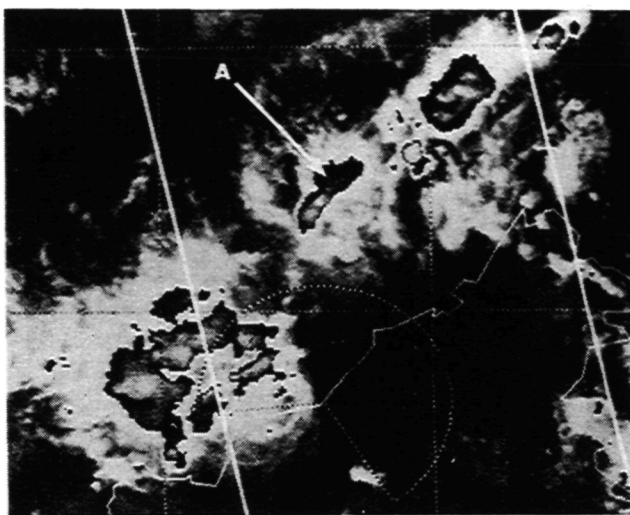
2–4. The original resolution of the visible and IR imagery was about 2 and 5 km, respectively; after remapping, the effective resolution is slightly coarser. Each set of images is ordered chronologically from top to bottom, so that the GMS-1 and SMMR data together provide a sequence of images covering a 3 h period. For the infrared images, we have used a “double ramp” gray-scale enhancement curve in order to highlight the brightness temperature structure in the coldest cloud tops. The “break” in the gray scale corresponds to a temperature of 213 K, and the whitest areas within this contour correspond to brightness temperatures as cold as 190 K. Such temperatures imply cloud top heights of roughly 13 and 17 km ($\approx$ tropopause height), respectively, based on the mean temperature profile for the WMONEX period.

a. 11 December

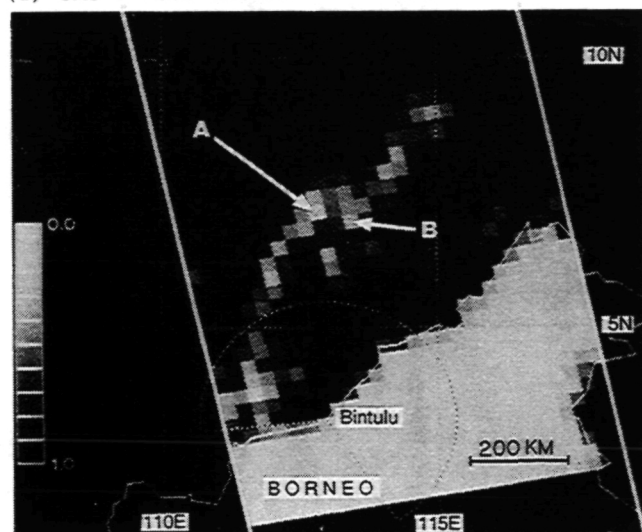
On 11 December 1978, at 0412 UTC (1212 LST) an extensive chain of convective cloud clusters over the South China Sea was sampled by the SMMR on Nimbus 7. The SMMR 37 GHz normalized polarization field P and the GMS-1 images for 0233 and 0533



(a) GMS-1 Visible 0233 UTC

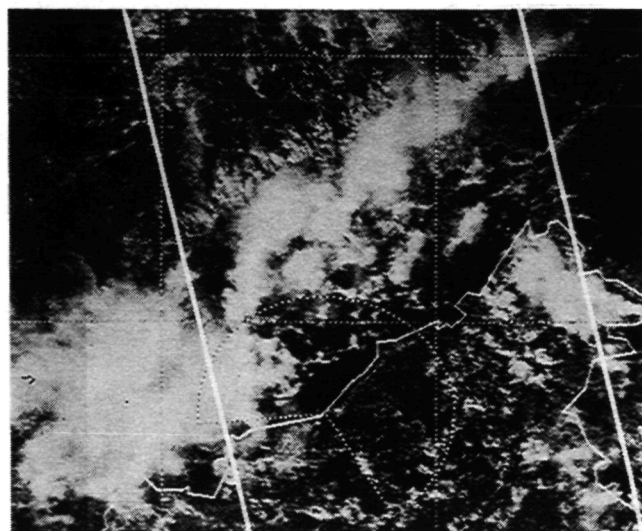


(b) GMS-1 Infrared 0233 UTC

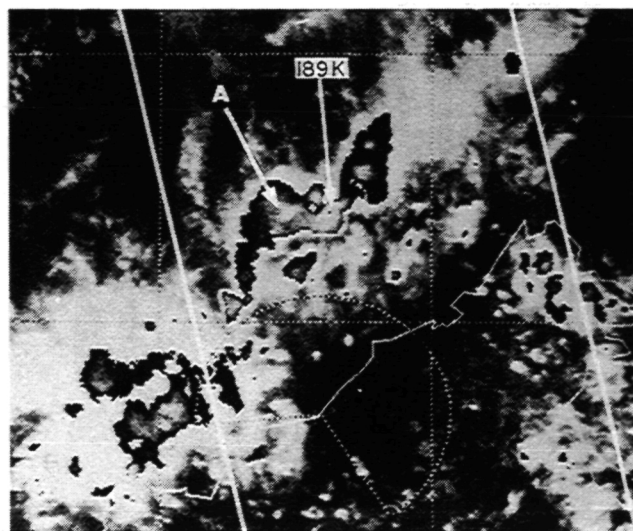


(c) SMMR: 0412 UTC

FIG. 2. SMMR and GMS-1 images of tropical cloud cluster over the South China Sea off the Borneo coast, 11 December 1978. Dashed semicircles in GMS-1 images indicate area of coverage by the WR-73 Bintulu radar. Solid white lines show edges of SMMR data swath. (a) GMS-1 visible image at 0233 UTC. (b) GMS-1 infrared imagery for same time and area as in (a). The gray-scale enhancement curve consists of a double ramp covering brightness temperatures from 310 K to 190 K, with the break occurring at 213 K (-60°C). (c) Normalized 37 GHz polarization difference (P) obtained from Nimbus-7 SMMR at 0412 UTC. Brightest pixels correspond to lowest polarization. (d) Same as in (a), but at 0533 UTC. (e) Same as in (b), but at 0533 UTC.

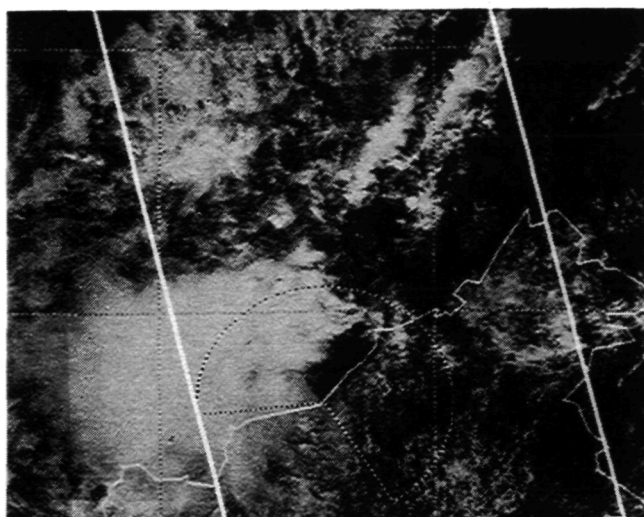


(d) GMS-1 Visible 0533 UTC

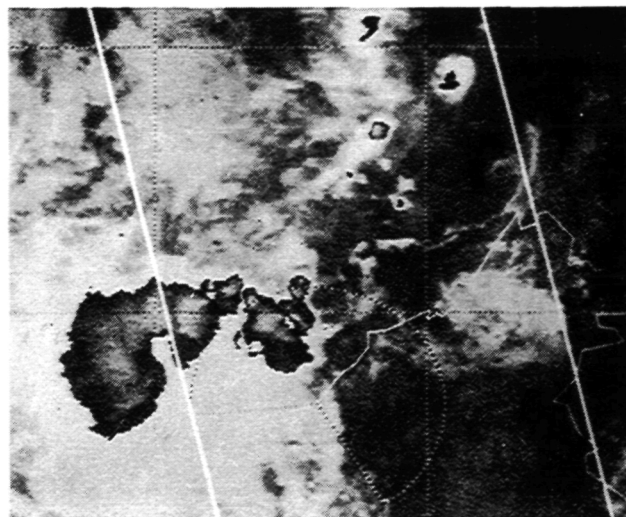


(e) GMS-1 Infrared 0533 UTC

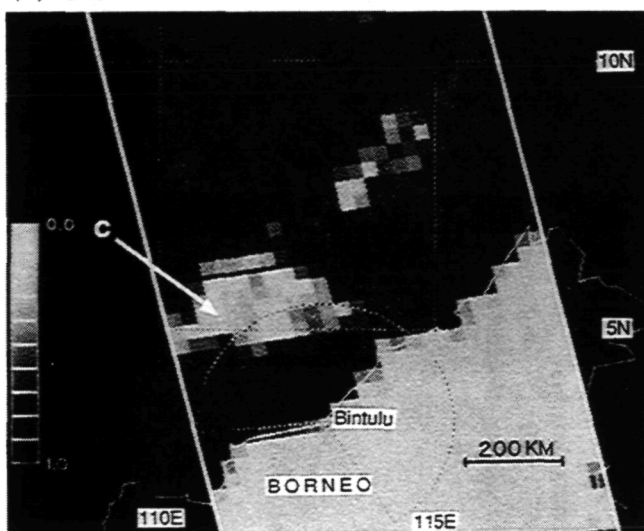
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(a) GMS-1 Visible 0233 UTC

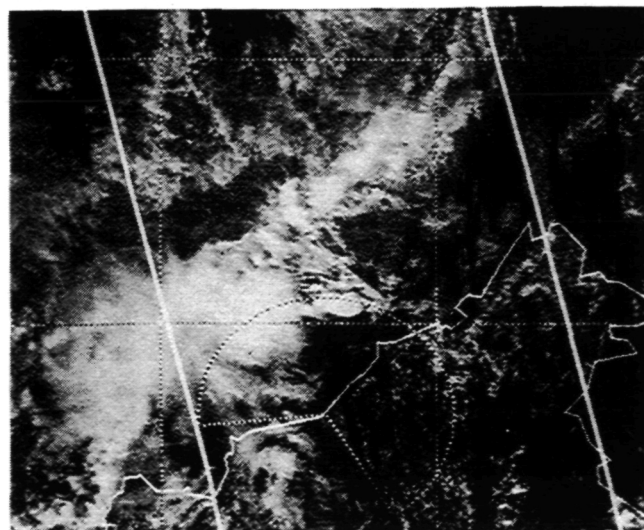


(b) GMS-1 Infrared 0233 UTC

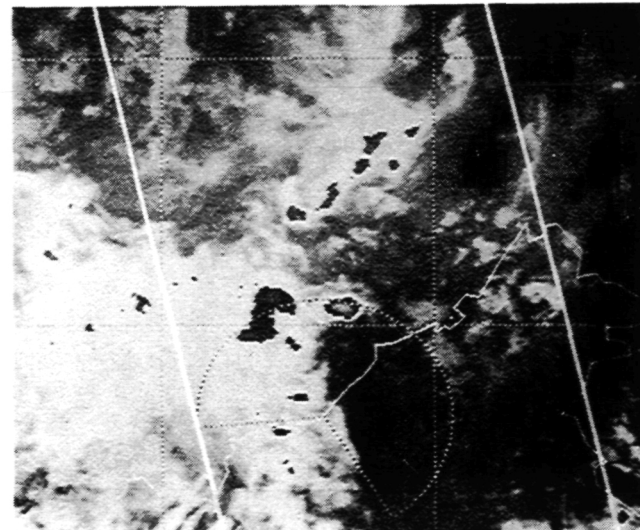


(c) SMMR: 0411 UTC

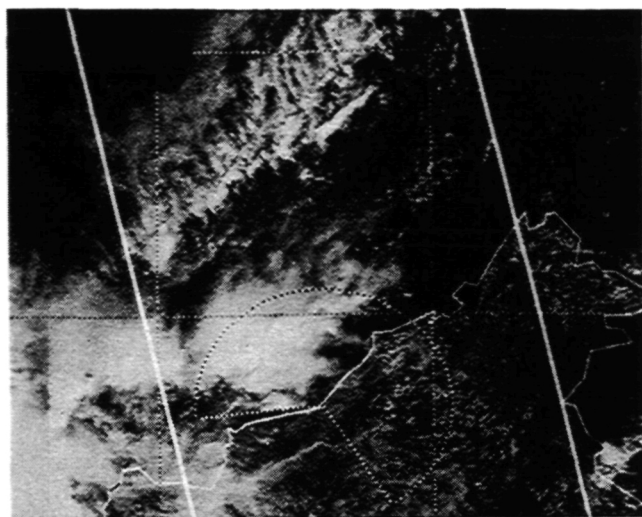
FIG. 3. As in Fig. 2 but for 17 December 1978. (a) GMS-1 visible image at 0233 UTC. (b) GMS-1 infrared image at 0233 UTC. (c) SMMR normalized polarization difference (P) at 0411 UTC. (d) GMS-1 visible image at 0533 UTC. (e) GMS-1 infrared image at 0533 UTC.



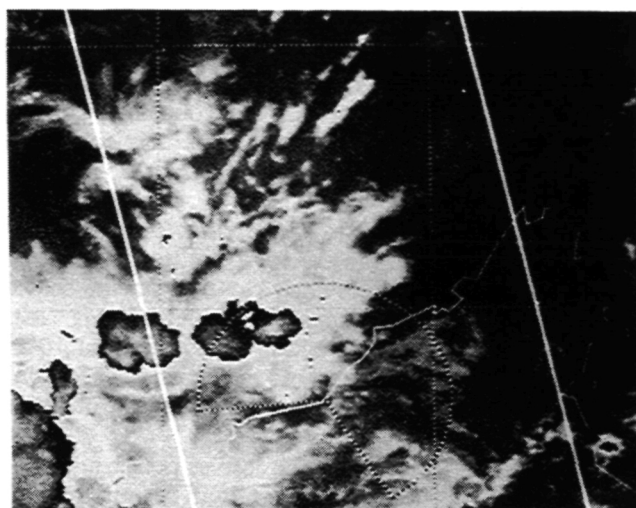
(d) GMS-1 Visible 0533 UTC



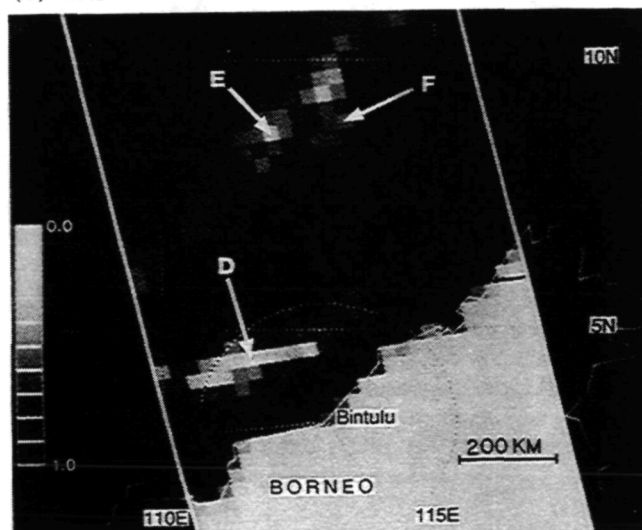
(e) GMS-1 Infrared 0533 UTC



(a) GMS-1 Visible 0233 UTC

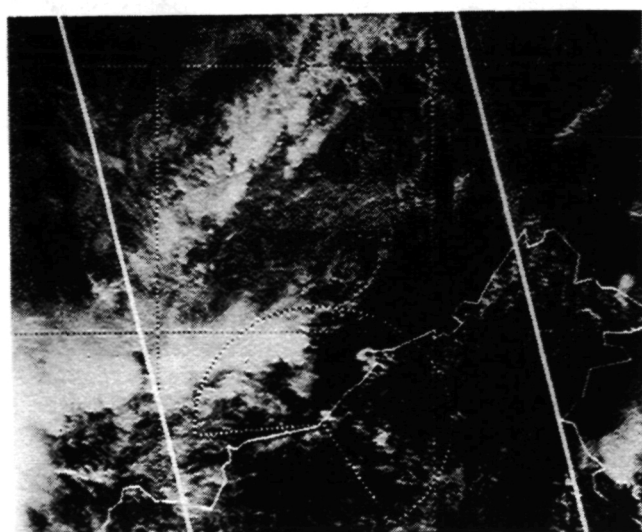


(b) GMS-1 Infrared 0233 UTC

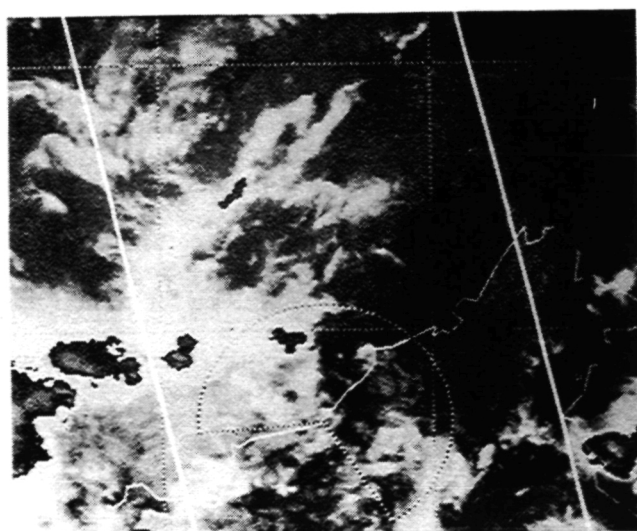


(c) SMMR: 0415 UTC

FIG. 4. As in Fig. 2 but for 23 December 1978. (a) GMS-1 visible image at 0233 UTC. (b) GMS-1 infrared image at 0233 UTC. (c) SMMR normalized polarization difference (P) at 0415 UTC. (d) GMS-1 visible image at 0533 UTC. (e) GMS-1 infrared image at 0533 UTC.



(d) GMS-1 Visible 0533 UTC



(e) GMS-1 Infrared 0533 UTC

UTC are shown in Figs. 2a–e. At the time of the SMMR pass, the intensity of the convective activity within radar range of Bintulu was decreasing, but further to the north, deep convection within the cloud mass continued to develop, as evidenced by extremely cold (189 K) cloud tops in the infrared image at 0533 UTC (Fig. 2e).

If the SMMR image is translated southwest by approximately 50 km, the general location and shape of the area of low P agrees fairly closely with the colder IR brightness temperatures at 0533 UTC (Fig. 2e), suggesting an overall advection speed of about 10 m s^{-1} . This corresponds closely to the observed speed and direction of the low-level monsoon flow throughout the period.

The SMMR pixel with the lowest polarization ($P = 0.15$) is labeled “A” in Fig. 2c. The horizontally polarized brightness temperature $T_{B,H}$ for this point is 233 K (see Fig. 1), several degrees colder than the expected $T_{B,H}$ for this value of P . Indeed, few other data points in Fig. 1, which includes SMMR data from all three cloud cluster cases, fall so far below the diagonal band defined by most of the remaining points. The combination of low polarization and depressed $T_{B,H}$ suggests that pixel A corresponds to a location of particularly strong convective activity. Low P , of course, implies that the pixel is nearly filled with optically thick rain cloud; the depressed $T_{B,H}$ is apparently due to the microwave scattering mechanism described by Spencer (1986), hence we may infer relatively heavy concentrations of larger ice particles in these clouds.

The approximate position of pixel A, adjusted for the apparent motion of the cluster, is also indicated in the IR images (Figs. 2b and 2e). From 0233 to 0533 UTC, the temperature of the cloud tops at A dropped from 210 K to about 193 K, and the area inside the 213 K contour grew to over 100 km in diameter. The GMS-1 images thus appear to confirm the presence of intense development at A through the time of the SMMR pass. At 0533 UTC, even colder (189 K), less diffuse cloud tops are evident about 100 km east (Fig. 2e) of A, and these appear to be associated with a relatively new center of very deep convection. At the time of the SMMR pass, about 1.4 hours earlier, development at this location had apparently not progressed to the stage where it could give rise to as strong a T_B depression as in pixel A; however, in its incipient phase, it may perhaps have been responsible for the fairly low polarization ($P = 0.31$) of pixel B.

The southern end of the cluster (near Bintulu) was observed by the MIT radar beginning at 0407 UTC, five minutes prior to the SMMR overpass. Discussion of radar–SMMR comparisons is deferred to section 5.

b. 17 December

On 17 December, a large, persistent cloud mass was observed in the GMS-1 image of the ocean area to the

north and west of Bintulu (Fig. 3). The main cluster appeared to weaken somewhat from 0233 to 0533 UTC, as indicated by the near disappearance of cloud top IR brightness temperatures below 213 K (Fig. 3b, 3e), but there were indications of strong new convection to the east of the old cloud shield, near 5.5°N , 113.5°E .

An extensive region of precipitation is observed in the same area by the SMMR overpass at 0411 UTC (Fig. 3c). Within the cloud mass, nine SMMR pixels exhibit values of $P < 0.15$ (i.e., below the lowest value of P observed in the 11 December cluster), suggesting at least moderate rain throughout the corresponding FOVs. The horizontally polarized 37 GHz T_B for these pixels ranges from about 241 to 254 K (Fig. 1), or about 20–30 K warmer than the IR brightness temperatures in the same locations. This difference clearly illustrates the ability of the SMMR to “see” the warmer levels of the atmosphere several kilometers below the tops of the thick anvil cirrus, in contrast to the infrared radiometer, which typically sees only the upper few tens of meters. Indeed, because of the “cooling” effect that radiative scattering has on the microwave T_B , the actual difference in effective emitting level for each wavelength is probably much greater than the $\sim 5 \text{ km}$ implied by the T_B difference alone.

Most striking in this case is the ability of the SMMR to delineate the region of precipitation beneath the extensive and rather amorphous cirrus shield seen in the GMS-1 IR and visible images. There appears to be no systematic relationship between the locations of the coldest IR cloud tops and the lowest P values, although this may be due, in part, to the 1.5 hour time difference. The same is true for pixels exhibiting markedly depressed T_B due to scattering (e.g., pixel C, Figs. 1 and 3c).

At the time of the overpass the radar showed no significant echoes to the north, despite the fact that the cloud cluster lay partially within its nominal 256 km range. The combined evidence of the GMS-1 images and the SMMR observations suggests that the radar coverage at this range and azimuth was deficient. This is consistent with the low-level beam blockage noted by Geotis and Houze (1985) for the Bintulu radar site (see section 5a).

c. 23 December

In the early morning of 23 December, an east–west oriented series of convective cloud clusters developed about 200 km to the northwest of Bintulu and continued to be visible in the GMS-1 images of 0233 and 0533 UTC and in the SMMR data swath at 0415 UTC (Fig. 4). There is an apparent overall weakening of the system between 0233 and 0533, as seen in the general warming of the cloud top temperatures. The weakening, and the accompanying trend from convective to stratiform precipitation, is also clearly indicated by the

corresponding series of radar observations (not shown). Thus, by the time of the SMMR overpass, this cloud cluster was well into its decaying phase, in keeping with the diurnal cycle documented by Houze et al. (1981b). The sequence of satellite images reveals no obvious systematic motion of the clusters, though this may again be simply due to the difficulty of identifying and tracking individual cloud features in images so widely separated in time.

As in the previous case, comparison of the SMMR and GMS-1 images reveals the SMMR's ability to delineate the region of rain within the more extensive cloud mass observed in both the IR and visible images. Overall, values of P here are higher than in the 11 December and 17 December cases, the minimum occurring at pixel D ($P = 0.22$). The relatively high minima are probably due to relatively low fractional coverage of the SMMR pixels by rain, since much of the main precipitation feature in Fig. 4c appears to be only one SMMR pixel wide (i.e., 30 km), and may in fact be considerably narrower than the SMMR can resolve.

As before, there is also clear relationship between the shapes and locations of IR $T_B < 231$ K and the locations of lowest P . Interestingly, a few of the more depressed polarizations (e.g., pixel E, $P = 0.33$) and the most strongly depressed microwave T_B (pixel F, see also Fig. 1) in this swath are associated not with the main cloud cluster but rather with some less prominent convective features near 8°N . Because we could not unambiguously track individual cloud elements over time from one image to the next, we have not attempted to identify the cloud features in the GMS-1 images which correspond to the labeled SMMR pixels.

The MIT radar at Bintulu recorded portions of the cluster beginning at 0410 UTC, five minutes prior to the SMMR overpass.

5. Results

a. Radar-SMMR comparisons

In Figs. 5a and 5b, the portion of the SMMR pass which fell within radar range of Bintulu is depicted for the 11 and 23 December cases. Overlaid on the SMMR normalized polarization field in each case is a map of the maximum radar reflectivity (dBZ) occurring below 4 km. For clarity, only two dBZ thresholds were used in displaying the echoes: the black regions represent the minimum detectable echo, the inner white regions represent reflectivities greater than 20 dBZ. The heavy white line identifies the region chosen for quantitative comparisons (see below). The difference between the horizontal scale of variability of precipitation and the spatial resolution of the SMMR becomes strikingly apparent in this figure.

Detailed contour plots of the radar reflectivities appear separately in Fig. 6. Reflectivity values displayed represent the maximum in the layer above the surface

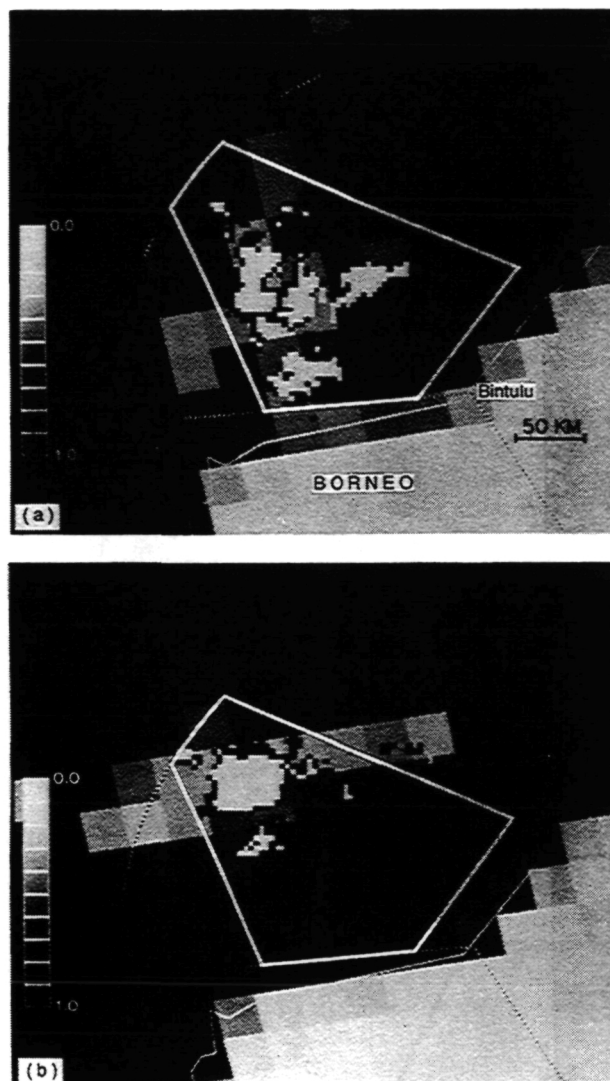


FIG. 5. Observed radar patterns overlaid on coincident SMMR polarization pattern. Dashed arc represents range of 256 km from radar site at Bintulu. Black radar pixels represent echoes exceeding the minimum detectable signal; white pixels represent values exceeding 20 dBZ. Solid white line borders area used for quantitative comparisons (see text). (a) Radar data from 0407 UTC; SMMR data from 0412 UTC, 11 December 1978. (b) Radar data from 0410 UTC; SMMR data from 0415 UTC 23 December 1978.

and below the bright band region. Whereas the 11 December case (Fig. 6a) appears as a patchy distribution of weak stratiform rain containing a few convective cores with radar reflectivities of up to 45 dBZ [46 mm h^{-1} , from Eq. (4)], the 23 December case (Fig. 6b) is best characterized as a relatively homogeneous region of stratiform rain, with maximum reflectivities of 38 dBZ (14 mm h^{-1}).

In Fig. 5, several SMMR pixels show significantly reduced polarization even in the absence of significant radar echoes. As already pointed out, the 17 December cloud cluster (Fig. 3) failed to appear in the radar data,

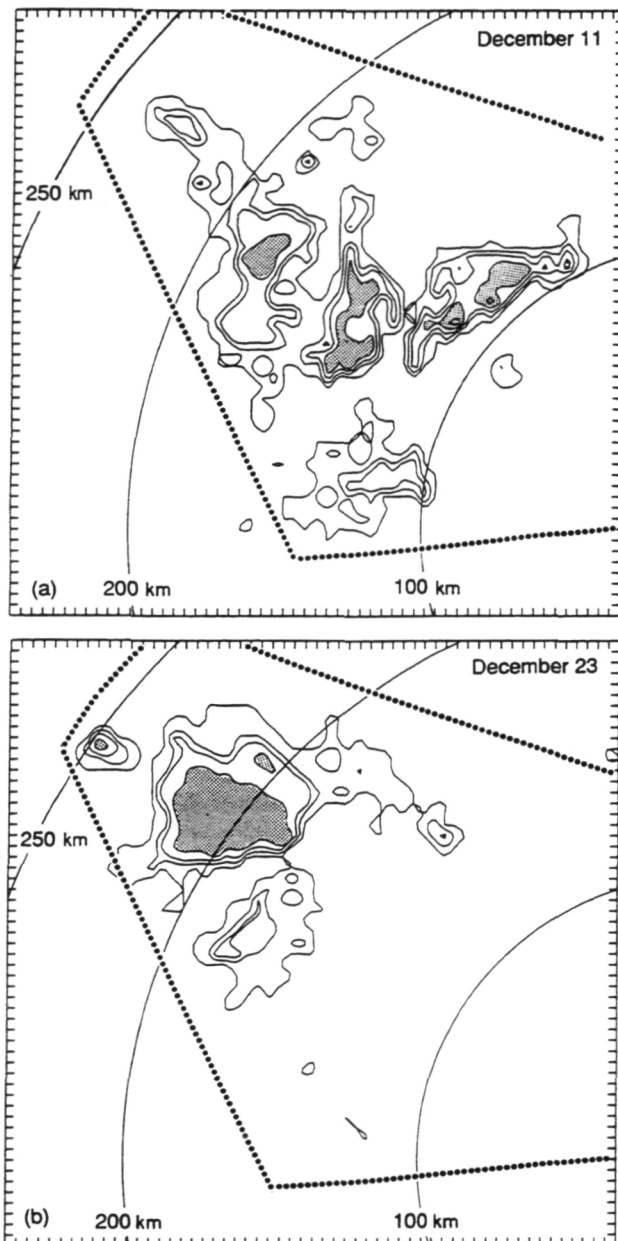


FIG. 6. Contours of dBZ recorded by MIT WR-73 radar. Contour values range from 0 dBZ to 40 dBZ, in 10 dBZ increments. The stippled areas represent reflectivity greater than 30 dBZ. Dotted line borders area used for quantitative comparisons (see text). Refer to Fig. 5 for geographical placement. The presence in this plot of reflectivity values less than the minimum detectable signal is an artifact of the interpolation to cartesian coordinates. (a) 0407 UTC 11 December 1978. (b) 0410 UTC 23 December 1978.

despite satellite evidence for its presence partially within the nominal range (256 km) of the radar. Geotis and Houze (1985) presented composited radar data for the entire 24 day period of WMONEX which showed that rainfall was rarely recorded beyond about 100 km range over water, except in that region roughly northwest of

Bintulu, i.e., aligned with the airport runway near which the radar was sited. At other azimuths, nearby trees and buildings appear to have significantly impaired the radar coverage. This observation led us to restrict quantitative SMMR-radar comparisons to SMMR pixels falling within the region indicated in Fig. 5, corresponding approximately to the outer boundary of the composited radar echo data shown by Geotis and Houze (1985). All of the obvious discrepancies between radar and SMMR, including the entire 17 December cluster, are found to fall outside of this area.

Radar pixels were divided among the overlapping SMMR pixels. The roughly 64 radar pixels thus assigned to each individual SMMR pixel were used to calculate several parameters of interest, including fractional coverage by echo exceeding the radar minimum detectable signal, and area-averaged rain rate using the $Z-R$ relation given by (4). No attempt was made to weight the radar values according to their position within a SMMR cell due to uncertainties in the navigation.

Figure 7 shows a scatter plot of P versus $\bar{R}$. Nonzero rain rates are seen to occur primarily in conjunction with values of $P < 0.8$. Beyond this, however, P appears to contain little information about the actual average rain rate. The correlation coefficient calculated for P and $\bar{R}$, (for $P < 0.9$ only) is unimpressive ($r = -0.48$).

A scatter plot (Fig. 8) of P versus fractional coverage by radar echo (F_E), however, clearly shows a systematic increase in F_E with decreasing P . For these two parameters, the calculated correlation is considerably improved ($r = -0.81$). Simple linear regression yields a

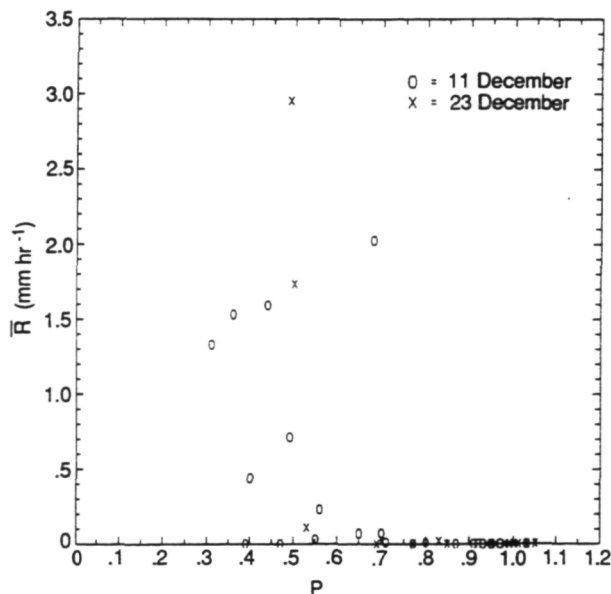


FIG. 7. SMMR 37 GHz normalized polarization difference (P) vs radar-derived average rain rate $\bar{R}$ for corresponding SMMR pixel. Total sample consists of 36 points.

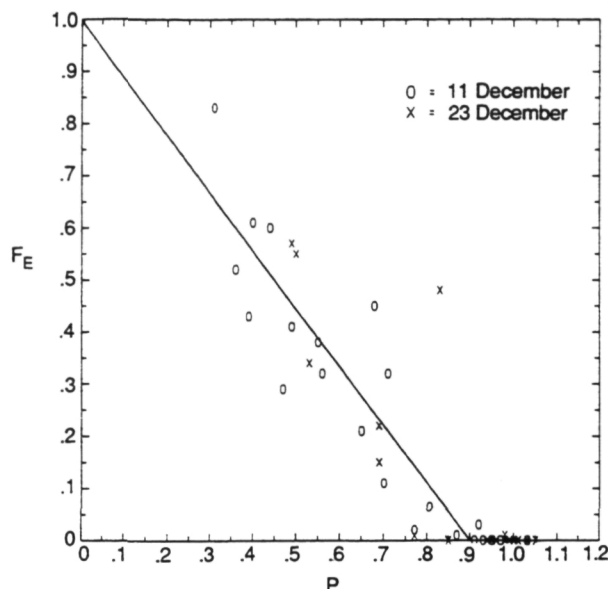


FIG. 8. SMMR 37 GHz normalized polarization difference (P) vs fractional radar echo coverage F_E for corresponding SMMR pixel. Total sample consists of 36 points. Diagonal line corresponds to Eq. (5) in the text.

best-fit line with a slope of -1.097 and an intercept of 0.99 . Because of the small number of data points ($n = 24$) and the degree of scatter about the fitted line, however, there is considerable statistical uncertainty in the regression-derived slope and intercept. With negligible loss of fit, therefore, we may round the slope and intercept to the nearby values $-1/0.9$ and 1.0 , respectively, so that the observed relationship may be expressed in the convenient form

$$F_E \approx \begin{cases} 1 - \frac{P}{0.9}, & \text{for } P < 0.9 \\ 0, & \text{for } P > 0.9 \end{cases}, \quad (5)$$

subject to refinement as more radar comparison data become available. This result is qualitatively consistent with (3) and reflects the fact that a large fraction of the area where it is raining consists of rain rates in excess of 3 mm h^{-1} . Some of the scatter in Fig. 8—particularly those points falling above the regression line—is undoubtedly due to the presence of some lighter rain rates as well, so that the assumption in (3) of optically thick rain does not strictly hold everywhere. The remaining scatter could be due to (i) residual navigation errors in the satellite data, (ii) our decision not to use a realistic spatial weighting function when calculating pixel statistics, (iii) errors in the radar data, or (iv) variations in the contribution of nonprecipitating cloud to the effective rain cloud opacity. The difference in the form of (3) and (5) can be readily explained by the ability of nonprecipitating cloud liquid water alone to account for values of P as low as 0.9 , a possibility which was neglected in (3).

In addition to the above comparisons, we also examined the results of the scattering-based 37 GHz method of Spencer (1986) compared with our radar data, using slightly modified coefficients to account for the different open-ocean 37 GHz T_B . Because of the small sample size available and the absence of a sufficiently large range of pixel-averaged rain rates within the comparison region, a meaningful correlation was not expected, nor was it found. We did observe, however, that the Spencer method tended to produce rain rate estimates which were rather noisy (standard deviation $\sim 10 \text{ mm h}^{-1}$), even outside of clouds. This result would appear to place an a priori limit on the minimum pixel-averaged rain rate that can be reliably detected with the scattering-based technique applied to the gridded SMMR "Cell-All" data. This appears, however, to be due to the high sensitivity of the Spencer (1986) method to noise that is introduced by the gridding process; our findings, therefore, cannot necessarily be interpreted as extending to his method applied to ungridded SMMR data (Spencer, personal communication 1989).

b. SMMR-derived precipitation coverage

Using (5) to estimate the fractional filling of SMMR pixels by rain, total precipitation coverage was calculated for portions of the mesoscale cloud clusters of 11, 17, and 23 December. The 11 December cluster (Fig. 1) was geographically the largest sampled by the SMMR, with rain occurring within a total of 108 of the SMMR pixels (each with an area of approximately 900 km^2). The average fractional coverage by rain within those pixels, however, was only 26%, giving rise to a total rain area of approximately $2.6 \times 10^4 \text{ km}^2$. This value, of course, does not account for the portion of the cluster to the west of 111°E , which fell outside the SMMR swath.

The 17 December cluster was also large, with 75 SMMR pixels affected by rain in the single broad cloud mass south of 7°N (Fig. 3c). The average fractional coverage in this case was significantly higher (49%), however, so that the total estimated rain area within the SMMR swath was larger ($3.3 \times 10^4 \text{ km}^2$) than observed in the 11 December cluster.

The portion of the 23 December cluster (south of 7°N , Fig. 4) which fell within the SMMR swath was the smallest in terms of both the number of pixels affected by rain (43) and total estimated rain area ($1.1 \times 10^4 \text{ km}^2$), although the average fractional coverage (29%) was slightly higher than in the 11 December case.

Because of evolution of the cloud fields between successive GMS-1 images, it was not possible to compare directly the SMMR estimates of precipitation area with the areas of the corresponding cirrus shields. Qualitatively, however, it appears that the combined areas of the SMMR pixels affected by precipitation were

considerably smaller than those of the cirrus shields depicted in the GMS-1 images. Since, in addition, the average fractional coverage within those pixels ranged from approximately one-quarter to one-half, it may be tentatively concluded that rain coverage in tropical cloud clusters at the stage of development observed here is much less than 50% and probably as low as 10%–20% of the area of the mesoscale cloud shield. For comparison, Churchill and Houze (1984) found a maximum ratio of radar-derived precipitation area to cirrus shield area (determined using a cloud top threshold temperature of 247 K) of about 20% in a detailed case study of another cloud cluster during WMONEX.

6. Conclusions

In this paper we documented the ability of polar-orbiting passive microwave radiometers to map the regions of precipitation in selected tropical cloud clusters. In each of the cases discussed, the SMMR shows evidence of being able to distinguish between major regions of rain and no rain, even beneath the cirrus shields accompanying the mesoscale cloud clusters.

We also sought to establish whether SMMR 37 GHz brightness temperature observations could be used to estimate area-averaged rain rate within tropical cloud clusters of the type encountered in the South China Sea. This was accomplished by correlating a microwave index of precipitation activity, the normalized polarization difference introduced in this study, with coincident radar estimates of FOV-averaged rain rate. The results of this comparison were disappointing, but not unexpected, in view of the strong spatial inhomogeneities inherent in tropical precipitation.

Nevertheless, we found that the SMMR 37 GHz normalized polarization difference is a rather good predictor of fractional coverage by rain ($r = -0.81$). A tentative quantification of the observed relationship is given by (5). Because (5) was derived from a small comparison dataset, further validation and refinement of this relationship for tropical oceanic rainfall is clearly needed.

Although our radar dataset was not suitable for rigorously testing the proposed scattering-based rain rate algorithm of Spencer (1986), due in part to the absence of sufficiently high FOV-averaged rain rates, we did find that the method applied to the gridded (Cell-All format) SMMR data yielded relatively noisy estimates even in nonraining situations. We therefore recommend caution in interpreting rain rate estimates falling much below, say, 20 mm h^{-1} when using the gridded SMMR T_B with this method. Direct validation of scattering-based algorithms is still needed for tropical oceanic convection, for the reasons discussed in section 2a.

We applied (5) to the SMMR observations of the cloud clusters in order to obtain estimates of total area coverage by precipitation within the SMMR swath, as well as average fractional coverage by precipitation

within the active regions of the clusters. These calculations represent perhaps the first direct satellite estimates of these parameters for precipitation over the near-equatorial oceans. An obvious application of this technique is the development of a global climatology of precipitation areas in tropical cloud clusters using microwave radiometers in space. Also, despite the apparent lack of skill of the SMMR 37 GHz polarization measurements at retrieving rain intensity on a pixel-by-pixel basis, we expect that the fractional coverage of rain obtained with these channels will prove to be usefully related to area-averaged rain rate for large enough samples.

These results have application to the goals of the proposed Tropical Rainfall Measurement Mission (TRMM, Simpson et al. 1988). As presently envisioned, this program will exploit simultaneous measurements by several microwave radiometer channels, with considerably improved surface resolution, in conjunction with the first spaceborne weather radar. This combination of instruments should prove to be an effective means to determine the three-dimensional structure and global distribution of tropical rain.

Our use of the normalized polarization difference to characterize rain clouds is subject to some of the same uncertainty as previously published relations, and this parameter in fact contains much of the same basic information as the more commonly used horizontally polarized brightness temperature. It departs from other methods in that it provides a useful estimate of a physically meaningful quantity related to rain rate, namely, the effective transmission of the polarized sea surface signal through and/or between the rain clouds, without resorting to numerical radiative transfer calculations, and it implicitly removes from the measurements much of the effect of variations in atmospheric water vapor content, surface wind roughness and sea surface temperature.

With this study, we learned that radar-derived precipitation patterns, though often treated as "ground truth" for validating satellite algorithms, are far from infallible. Indeed, we found that all of the more obvious discrepancies between SMMR and radar could be readily explained by deficiencies in the radar coverage which were documented only many years after the WMONEX project. It is difficult to say how much of the residual scatter in the plots of radar versus SMMR is in fact due to undetected errors in the radar data. Since there is no reason to think that such problems are entirely unique to the WMONEX radar site, we recommend that future comparisons between radar and satellite data be conducted with great care. On the positive side, it appears that satellite microwave data should prove useful for identifying similar defects at other coastal radar sites.

A new microwave radiometer, the Special Sensor Microwave Imager (SSM/I), now operational on board a satellite in the Defense Meteorological Satellite Program (DMSP), will permit continued development and

application of emission-, attenuation-, and scattering-based passive microwave techniques pioneered with the SMMR and earlier sensors. In addition to several basic design improvements, the SSM/I incorporates new channels at 85.5 GHz which, because of their relatively high resolution (12 km) and strong sensitivity to rain and precipitation-size ice particles, appear particularly well suited to the task of observing subsynoptic scale distributions of both midlatitude and tropical precipitation.

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***Nimbus-7* SMMR Precipitation Observations Calibrated against Surface Radar during TAMEX**

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ABSTRACT

Nimbus-7 Scanning Multichannel Microwave Radiometer (SMMR) normalized 37-GHz polarization differences " P " were compared with surface digital radar observations of oceanic precipitation made during the Taiwan Area Mesoscale Experiment (TAMEX). Four cases were found for which SMMR and radar coverage of significant precipitation features were nearly simultaneous. These yielded 518 SMMR-radar data pairs, of which over half included precipitation. An empirical method was used to correct the radar data for range-dependent errors, and relationships were then sought between the corrected pixel-averaged radar rainfall parameters and the SMMR-observed microwave polarization P . Because of the small sample size and large statistical uncertainties associated with the direct comparisons, the latter were most useful as a means to validate and tune a theoretically derived relationship between 37-GHz P and radar reflectivity factor Z . This relationship in turn was used to generate a large set of simulated SMMR observations from all available TAMEX radar scans in order to produce histograms and mean values of pixel-averaged rain rate as a function of P . An Appendix also describes an attempt to compare SMMR estimates of integrated cloud liquid water with coincident aircraft data during TAMEX.

1. Introduction

In a recent study, Petty and Katsaros (1990a, hereafter referred to as PK) examined satellite observations of tropical cloud clusters over the South China Sea using the 37-GHz channels of the *Nimbus-7* Scanning Multichannel Microwave Radiometer (SMMR). A normalized form of the 37-GHz polarization difference was introduced and compared with satellite visible and infrared images and nearly simultaneous surface digital radar observations of precipitation. The normalized polarization P was found in that study to be a good indicator of the location of precipitation within tropical cloud masses; however, values of P appeared to be much more strongly correlated with fractional coverage by rain than with area-averaged rain rate.

The primary aim of PK was to help build an empirical basis for the qualitative and quantitative interpretation of microwave radiometric observations of rainfall in the tropics. Previous comparisons between simultaneous surface measurements of precipitation and satellite microwave brightness temperatures had

been limited mainly to latitudes north of 25°N (e.g., Spencer et al. 1983; Olson 1989).

Aside from these few empirical studies, the principal basis for rain-rate retrieval algorithm development efforts, both for the tropics and for higher latitudes, has been theoretical radiative transfer modeling. Such models have been described by Wilheit et al. (1977), Jung (1980), Chang and Milman (1982), Wu and Weinman (1984), Kummerow and Weinman (1988), Mugnai and Smith (1988), Smith and Mugnai (1988), and Olson (1989). Unfortunately, the quantitative predictions of theoretical models are sensitive to the details of the rain-cloud microphysical and mesoscale structure assumed in the model. Many of these properties are still poorly known for precipitating systems over the tropical and subtropical oceans where in situ measurements are scarce.

In view of the planned Tropical Rainfall Measuring Mission (TRMM), which will depend heavily on both passive and active microwave sensing systems to establish an accurate rainfall climatology between approximately 30°N and 30°S (Simpson et al. 1988), it is important to exploit every opportunity to validate and refine our knowledge of the microwave signatures of precipitating cloud systems with the aid of in situ comparison data obtained from this latitude belt.

In this study, we seek to expand on the results of PK by examining the relationship of SMMR 37-GHz normalized polarization differences to surface digital radar observations of oceanic precipitation near the

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east coast of Taiwan. These observations were made during the Taiwan Area Mesoscale Experiment (TAMEX). The comparison dataset available from TAMEX is in many ways superior to the dataset employed in the earlier study, which examined coincident SMMR and radar observations over a rather restricted area of only two near-equatorial cloud clusters during the Winter Monsoon Experiment (WMONEX). The TAMEX dataset, because of its size and quality, lends itself to a more thorough analysis. Perhaps partly for this reason, the results and conclusions of the present study differ in several respects from those reported by PK.

Because dual-polarization 37-GHz channels are common to all recent and planned satellite-borne microwave imagers, including the current Special Sensor Microwave/Imager (SSM/I) and the planned TRMM microwave radiometer, the following results should be applicable to those sensors as well, after allowance for minor differences in incidence angle and spatial resolution.

2. TAMEX

The Taiwan Area Mesoscale Experiment (TAMEX) was conducted jointly by scientists from the Republic of China and the United States during May and June of 1987. One goal of this field experiment was to study precipitation events associated with the so-called Mei-Yu (or baiu) front, a characteristic synoptic feature that often brings heavy convective rainfall to Japan and Taiwan during the late spring and early summer. For an overview of the meteorology of the Mei-Yu, the observing strategy employed in TAMEX, and some preliminary results, the reader is referred to Kuo and Chen (1990).

For the purposes of the present discussion, it is sufficient to note that a number of Doppler and conventional radar sites were in operation along the Taiwan coast. Two of the conventional sites, those located at Kaohsiung on the southwest coast and at Hualien on the central east coast, were equipped to record digital reflectivity patterns over the ocean out to a nominal range of 500 km. Only the Hualien radar site sampled a useful number of precipitation events that coincided with SMMR overpasses during the TAMEX; therefore, the attention of this study is restricted to the Hualien dataset.

In addition to surface observing facilities, an instrumented P-3 research aircraft and crew were provided by the National Oceanic and Atmospheric Administration (NOAA) to make airborne Doppler and conventional radar observations of mesoscale precipitation features, as well as in situ observations of cloud microphysics, vertical velocities, and other meteorological variables. One of us manned the conventional radar station aboard the P-3 throughout the TAMEX field phase and was therefore able to observe firsthand many

of the physical characteristics of the storms studied in this experiment, including two storms that were viewed nearly simultaneously by the SMMR. Some of these general observations prove relevant to the interpretation of the results presented in the following sections.

One important motive for participation in TAMEX was interest in obtaining aircraft measurements of cloud liquid water, vertical and horizontal reflectivity structure, and precipitation microphysics that could be used to quantitatively evaluate simultaneous column liquid water retrieval algorithms and microwave radiative transfer calculations for the SMMR data. Unfortunately, because of the complex and rapidly evolving spatial structure of the cloud systems, this goal remained essentially unfulfilled, as it proved nearly impossible to obtain in situ data at the moment of satellite overpass, which could be compared in a meaningful way to the coarsely sampled satellite microwave data. An exception occurred on 17 June 1987 when a pair of coincident aircraft and SMMR estimates of vertically integrated liquid water were obtained. This case is discussed in the Appendix.

3. Radar data

a. Characteristics and processing

In at least one respect, the dataset derived from the Hualien WSR-74S radar during TAMEX is better suited to quantitative comparison with SMMR data than was the WMONEX dataset employed by PK. Unlike the radar at Bintulu in WMONEX, which was partially obstructed by trees and buildings except along the length of the airport runway, the radar antenna at Hualien is located in a small tower on a high embankment directly overlooking the ocean and has an unobstructed view clockwise from approximately north-northeast to due south. In inland directions the radar beam is blocked at close range by the central mountain range.

The physical characteristics of the radar are given in Table 1. Volume scans, each derived from 19 elevation scans and requiring a total of approximately 10 min time, were interpolated on-site to a three-dimensional Cartesian grid and recorded on digital tape. Only the Cartesian gridded data were retained. Horizontal grid spacing was 4 km, and vertical grid spacing was 1 km, but typically only a few selected levels were archived. In addition to these constant altitude reflectivity

TABLE 1. Characteristics of WSR-74S digital radar located at Hualien, Taiwan.

Wavelength	10.35 cm
Frequency	2700 MHz
Peak power	500 KW
Pulse repetition frequency	539 s ⁻¹
Antenna diameter	3.66 m
Beamwidth	2.25°

patterns, a column-vector (CV) reflectivity grid was recorded containing the maximum reflectivity values within each vertical column in the Cartesian grid. It is the CV data that is employed in the radar-SMMR comparisons discussed below.

Gridded echo intensities exceeding 0 dBZ were stored at a resolution of 0.5 dBZ. Since the true minimum detectable signal is well above 0 dBZ at most ranges, the very low (dBZ < 10) values that are found in the data at all ranges presumably arise as a consequence of the weighted interpolation of the radial reflectivities to Cartesian coordinates.

Conversion of radar reflectivity to rain rate was accomplished using a Z - R relationship appropriate for "average or ordinary rain" according to Austin (1987):

$$Z = 230 R^{1.4}. \quad (1)$$

(Recall that $\text{dBZ} = 10 \log_{10} Z$.) The choice of (1) was based on the observation that reflectivity data used in the comparisons corresponded to a mixture of cellular and widespread rain.

b. Limitations and corrections

When deriving and using empirical relations between satellite observations and radar-derived surface rainfall, it is crucial to understand and allow for the limitations of radar as "ground truth." Some well-known sources of error, such as beam spreading and overshooting (arising from finite antenna resolution and the earth's curvature, respectively), brightband contamination, atmospheric attenuation, and uncertainties in the true Z - R relationship (due to variations in drop-size distribution), are common to all radar sites. Others, including local or distant obstructions to the radar beam, radar calibration, or details of the sampling, processing, and interpolation procedures, may be site specific. An excellent overview of many of these problems is given by Joss and Waldvogel (1990).

Furthermore, all radars have a noise threshold below which reflected power cannot be detected. Hence, many weaker echoes may not be recorded at all, particularly at greater ranges. Of course, the problem of echo truncation is exacerbated by the loss or spatial diffusion of echo power accompanying many of the other error sources described.

In short, there are many difficulties with estimation of rain parameters from radar data, particularly at ranges in excess of ~ 100 km. Unfortunately, for the purpose of calibrating satellite microwave estimates of precipitation, there is the conflicting requirement that the comparison data be sufficiently distant from the coast to avoid land contamination of the microwave signal and that the comparison area be large enough to ensure an adequate sample of precipitation observed simultaneously by radar and the satellite sensor during the period in question.

We must therefore carefully consider the impact of range-related errors on the radar-derived rain parameters used in this study, and attempt to correct them where possible. Because so many physical factors are involved, each of which individually can only be crudely estimated given the available meteorological information, a purely theoretical correction scheme attempting to account for all of these would be difficult and of doubtful value. Instead, the following empirical approach has been developed, which is believed to significantly improve the quantitative usefulness of the radar dataset without being unnecessarily complex.

1) RANGE-DEPENDENT CORRECTIONS

The entire set of Hualien radar scans archived for TAMEX were considered. After screening for obviously bad data, this set consisted of 390 scans spanning the period 13 May–20 June 1987, minus several periods ranging from hours to days that were not archived because of the absence of echo activity. During active periods, the interval between archived volume scans ranged from 20 min to 1 h.

First, a single composite reflectivity pattern was produced that represented the entire TAMEX period and consisted of the highest reflectivity (dBZ) value observed at each horizontal grid point (Fig. 1). The results reveal a sharply defined sector of unobstructed coverage over the ocean, consistent with the expected influence of the nearby central mountain range. On the basis of this composite radar image, attention in the remaining discussion is restricted to reflectivities observed between azimuths 30° and 190° (relative to true north).

Within this azimuth sector, the composite reflectivity pattern is nearly radially symmetric about Hualien and exhibits a distinct decrease of maximum observed reflectivity with increasing range. There is little evidence of any other systematic spatial patterns in the distribution of maximum reflectivities observed.

Ideally, surface-based raingages would be used to detect and to empirically correct biased radar rainfall estimates. Unfortunately, the proximity of the central mountain range limits unobstructed radar coverage over land to a very narrow coastal strip within approximately 100 km of the site. However, it is precisely the data beyond this range that are susceptible to the most serious errors.

In the absence of either suitable raingage data or independent information concerning the climatological distribution of rainfall off the east coast of Taiwan, it has been assumed in the following that rain events were in fact more or less uniformly distributed throughout the sector during the period in question and that all of the apparent range bias reflects only the physical limitations of the radar site.

Therefore, the range-dependent statistics were computed from the complete set of 390 radar scans, integrated over azimuth within the usable sector. The solid

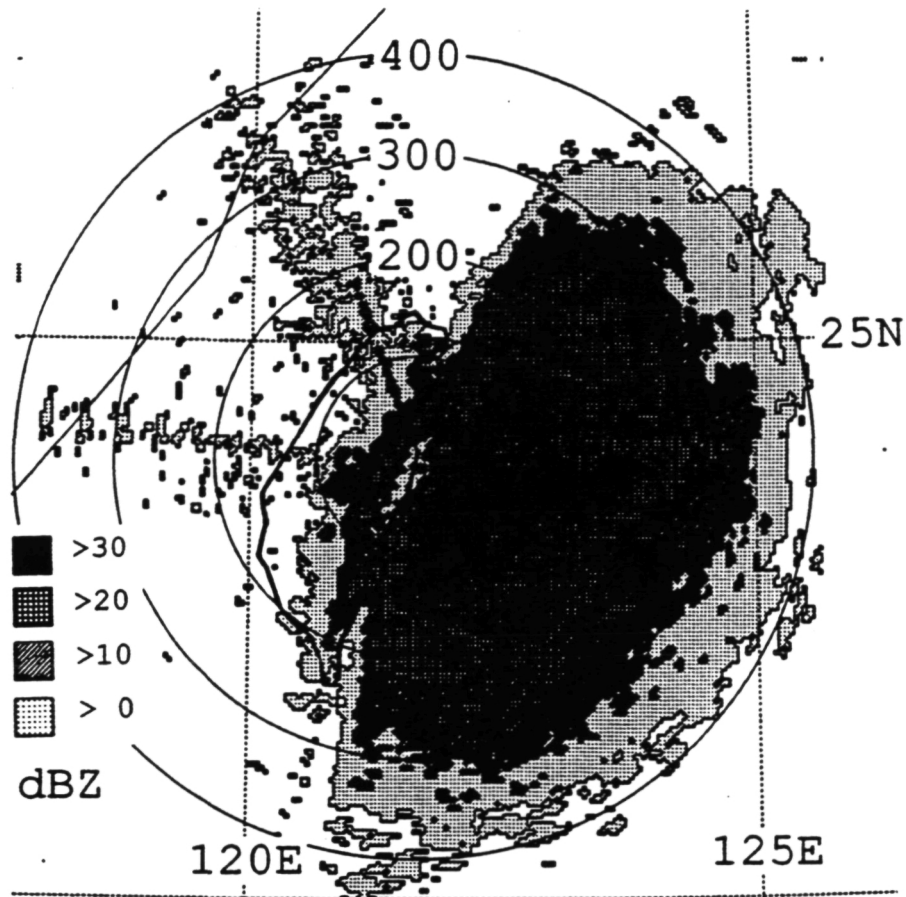


FIG. 1. Maximum radar reflectivity recorded at each grid point by Hualien radar during TAMEX period (13 May–18 June 1987). Rings indicate radar range at 100-km intervals.

lines in Figs. 2a–c depict the mean fractional coverage by echo observed within each range interval of 20 km, the radar-derived mean rain rate within echoes exceeding 17 dBZ, and the overall average rain rate, respectively. (Neither Fig. 2a nor Fig. 2c should be interpreted as true temporal averages for TAMEX, since the sample of radar scans included in the archive is biased in favor of the more meteorologically active periods.) Statistics within the interval 0–40-km range were discarded because of obvious contamination by surface clutter and because the statistics are derived from a small sample area.

As shown in Fig. 2, radar estimates of all three rainfall statistics are found to decrease dramatically with distance from Hualien. For instance, between 80 and 100 km, the observed mean fractional occurrence of echo is 30% and the average radar-derived rain rate within echoes exceeding 17 dBZ is 1.26 mm h^{-1} . Between 180 and 200 km, the mean fractional coverage falls to 17% and the average rain rate within echoes $> 17 \text{ dBZ}$ falls to 0.80 mm h^{-1} . The combined effect of the loss of both mean echo area and strength is an apparent decrease in average total rainfall from 0.11 to 0.026 mm h^{-1} , or more than a fourfold reduction.

Figure 3 (solid curves) shows the observed mean fractional coverage by echoes exceeding various reflectivity thresholds plotted as a function of range. Under the assumption that the true echo-area distribution is approximately range independent and that the apparent decrease with range is in fact the combined result of the many sources of error discussed in the previous subsection, the following correction was applied to all nonzero reflectivities at ranges d greater than 70 km:

$$\text{dBZ}' = \text{dBZ} + 0.075(d - 70); \quad d > 70. \quad (2)$$

Then the echo-area distributions for the Hualien radar dataset were recalculated using the adjusted reflectivities. With this simple linear correction, the echo distribution is remarkably flat out to about 280 km (Fig. 3, dashed curves), except for reflectivity thresholds $< 0.075(d - 70)$ whose correction would require knowledge of echoes whose uncorrected intensity was below the minimum recorded level of 0 dBZ.

Beyond 280 km, a surface-based radar in a standard atmosphere is unable to observe the atmosphere below $\sim 4.5 \text{ km}$, which approximately corresponds to the mean freezing level in TAMEX, that is, the approximate upper boundary of stratiform liquid precipitation.

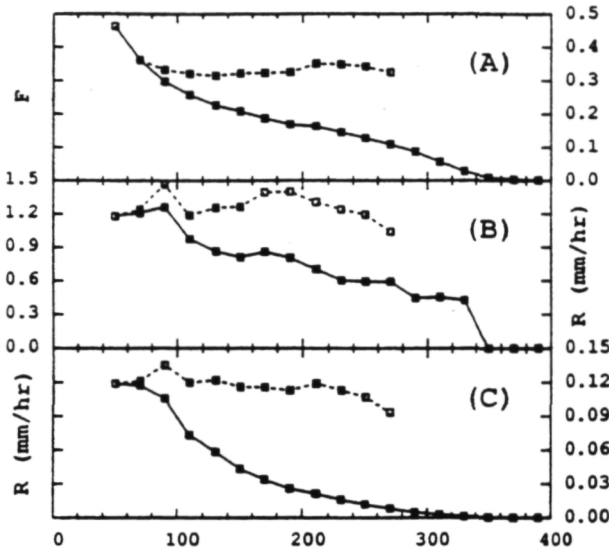


FIG. 2. Hualien radar rain statistics as function of range for the entire TAMEX period. Statistics were computed from 20-km range bins within the azimuth sector 30° – 190° relative to true north. Solid lines represent values obtained from the uncorrected radar data; dashed lines are from data corrected using (2)–(5). (a) Average fractional coverage by echo F ; (b) average rain rate within echoes exceeding 17 dBZ; (c) overall average rain rate.

This fact may explain the greatly accelerated falloff of fractional echo coverage beyond this range. It is therefore also unlikely that useful quantitative information can be recovered beyond this range, even with corrections.

Since unbiased estimates of area-averaged rain rate are required for comparison with SMMR observations, it is still necessary to account for the decrease with range of total echo area. The correction for lost echoes was premised on the following assumptions or observations:

(i) The true climatological average echo coverage is approximately independent of range, irrespective of reflectivity threshold.

(ii) Lost echo area arises exclusively from radar pixels whose uncorrected reflectivity value would have been less than 0 dBZ and which were therefore not recorded. In other words, all lost echoes represent combinations of range and uncorrected reflectivity value falling above the 0-dBZ contour in Fig. 3.

(iii) The mean loss of echo as a function of range and reflectivity threshold is thus given, in effect, by the vertical distance between the 0-dBZ contour in Fig. 3 and a hypothetical horizontal line representing the assumed true fractional coverage for the threshold in question.

(iv) It is impossible to know from which locations the echoes were lost; therefore, the correction scheme should distribute the presumed echo-area loss for a given range equally over all echo-free pixels at that range.

The range-independent true fractional coverage postulated in (i) and exploited in (iii) was modeled using a power-law fit to the corrected data within 150 km of Hualien:

$$\log_e(F_0) = -1.08Z^{0.218}, \quad (3)$$

where F_0 is the mean fractional coverage by echo exceeding Z in intensity. The fit of (3) to the observed mean fractional coverage (Fig. 4) is excellent for all reflectivity values that contribute significantly to the total mean rainfall. Note that reflectivities much greater than 30 dBZ were rarely observed, constituting less than 1% of the total observed echo area.

The position of the 0-dBZ contour in Fig. 3 (for $d < 280$ km) was then approximated with

$$f_i(d) = 0.49e^{-0.0055d}. \quad (4)$$

Following assumption (iv), the following formula was thus used to supply the estimated fractional echo coverage F'_Z at radar grid points for which no echo was recorded:

$$F'_Z = \frac{F_0(Z) - f_i(d)}{1 - f_i(d)}. \quad (5)$$

This correction is applied only at those ranges d for which $F'_Z > 0$.

Using (2)–(5), rain statistics were recalculated as function of range out to 280 km and are displayed as dashed lines in Fig. 2. All three parameters—mean fractional rain coverage, mean echo-averaged rain rate, and overall mean rain rate—are now almost constant with range.

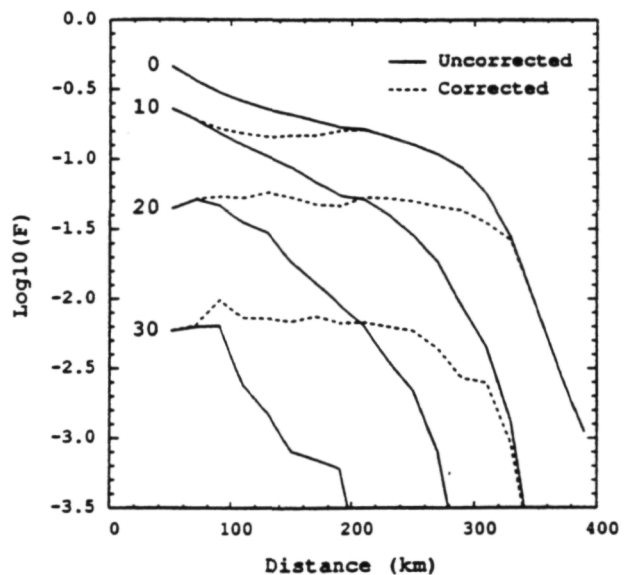


FIG. 3. Mean fractional coverage by echo as a function of range for reflectivity thresholds of 0, 10, 20, and 30 dBZ before correction (solid curves) and after correction using (2) (dashed curves).

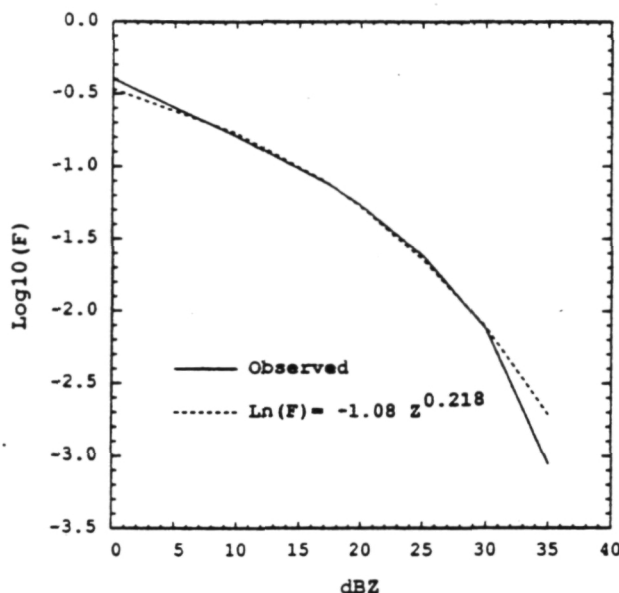


FIG. 4. Mean fractional coverage by echo as function of reflectivity threshold (corrected only) for the range interval 50–150 km. Solid line represents observed distribution; dashed line represents the analytic approximation given by (3).

It is worth emphasizing that, while the fractional echo-coverage correction represented by (3)–(5) can be quite substantial, only the very weakest echoes (e.g., rain rates $< 0.2 \text{ mm h}^{-1}$ at the maximum range of 280 km) are assumed to be lost. As a result, the consequences of the echo loss for the satellite–radar comparisons are not as severe as they might initially seem.

2) VERIFICATION OF ABSOLUTE CALIBRATION

Having thus accounted for apparent range-dependent errors, range-independent biases were accounted for by comparing Hualien radar reflectivities with reflectivities recorded by the lower fuselage radar on board the NOAA-42 P-3 aircraft. This radar, described in detail by Jorgensen (1984), was calibrated both immediately prior to and following the TAMEX field phase.

On 5 June 1987, a mesoscale precipitation feature located about 150–200 km south of Hualien was observed nearly simultaneously by both the Hualien radar and the P-3 lower fuselage radar. The most suitable aircraft radar data for comparison was recorded at 1506 UTC, at a range of 40–120 km, while the aircraft was in level flight near an altitude of 2000 m. The corresponding Hualien radar volume scan began at 1500 UTC.

Distributions of echo area as functions of reflectivity threshold were calculated for the P-3 instantaneous radar pattern and for the Hualien radar pattern (range corrected as described above) within identical areas of approximately $2 \times 10^4 \text{ km}^2$. The two distributions were

found to agree to within approximately 1 dBZ for almost all area-coverage thresholds. Area-averaged rain rate and echo-averaged rain rate (dBZ > 17 only) computed from the Hualien data were 0.37 and 1.20 mm h^{-1} , respectively, compared with 0.30 and 0.93 mm h^{-1} , respectively, from the P-3 lower fuselage radar.

The mean distance of the echoes from the aircraft was 82 km, at which range beam-filling problems begin to become important because of the 4.1° vertical beamwidth of the P-3 radar antenna. Therefore, the roughly 20% difference between the surface and aircraft radar rain estimates may be regarded as quite reasonable, and no basis can be found for questioning the absolute calibration of the Hualien data.

4. SMMR data and algorithm

Gridded (“Cell-All”) data from the *Nimbus-7* SMMR include, among other parameters, horizontally and vertically polarized 37-GHz brightness temperatures at approximately 30-km resolution. Each data swath is 26 pixels wide or about 840 km. The *Nimbus-7* satellite was in a polar sun-synchronous orbit, crossing the equator near noon and midnight local time.

For a more detailed description of the SMMR instrument and the motivation for using microwave brightness temperatures to observe precipitation from space, the reader is referred to PK and papers cited therein. The following is a brief review of the 37-GHz algorithm introduced by PK and used in the present study.

In order to permit a relatively simple and uniform physical interpretation of SMMR 37-GHz oceanic brightness temperatures T_B under varying conditions, PK proposed dividing observed polarization differences $\Delta T_B = (T_{B,V} - T_{B,H})$ by the hypothetical polarization difference $\Delta T_{B,Cl}$ expected in the absence of cloud and rain under otherwise similar conditions.

This procedure yields a nondimensional polarization difference P , which ranges from approximately unity in the absence of clouds to near zero in the presence of completely opaque rain clouds and which maintains a relatively invariant relationship to effective rain-cloud opacity, irrespective of variations in background polarization. The latter can arise from variations in wind-induced roughening of the ocean surface, atmospheric and sea surface temperature and/or atmospheric water vapor content. In summary, P is ostensibly a measure of the atmospheric opacity at 37 GHz due to liquid water alone, with all other sources of variability in the polarization difference having been largely factored out.

Determination of $\Delta T_{B,Cl}$ was accomplished using a histogram technique, the details of which are given by PK. Two minor modifications of the PK technique were incorporated to allow for the large fraction of the SMMR swaths affected by land in some of the TAMEX cases.

TABLE 2. Dates and times of radar-SMMR comparison cases.

Date	Radar time (UTC)	SMMR time (UTC)
13 May 1987	1530	1535
20 May 1987	0300	0311
5 Jun 1987	1540	1539
17 Jun 1987	1600	1551

The first of these is that the $(M \times M)$ -pixel window, which is searched for open-ocean pixels, was allowed to expand incrementally from $M = 13$ up to $M = 26$, as necessary, to ensure a sufficiently large sample of pixels unaffected by either heavy cloud or land. The second is that $\Delta T_{37, \text{Clr}}$ was constrained to be at least 40 K. This latter modification was designed to limit the errors that could arise in occasional cases for which the objective technique failed for lack of clear-sky pixels in the search window.¹

Later in this paper it becomes necessary to distinguish between the spatially averaged value of P observed by the SMMR and the higher resolution values of normalized polarization associated, say, with individual radar grid points. Henceforth, an overbar will therefore be employed to distinguish all FOV-averaged variables so that the SMMR-observed normalized polarization will appear as $\bar{P}$.

5. Cases

A search of the radar record and the available SMMR data for the TAMEX period yielded four significant precipitation events that were observed nearly simultaneously by the SMMR and the Hualien radar. The dates and times of these are listed in Table 2. Figures 5–8 show the SMMR $\bar{P}$ fields side-by-side with the corresponding observed radar reflectivity patterns (uncorrected).

The 13 May case occurred during the first TAMEX intensive observing period (IOP). A weak surface cold front moved rapidly from southeast China over Taiwan and triggered a narrow northeast–southwest-oriented band of precipitation east of Taiwan. At the time of the SMMR overpass (1535 UTC), the band consisted of a narrow line of relatively weak convective precip-

itation nearest the Taiwan coast south of Hualien and a mixture of convective and stratiform rain extending eastward out of the radar area (Fig. 5). The SMMR $\bar{P}$ field suggests that the most significant precipitation activity occurred outside the 200-km range marker, but the radar picks up only sparse, weak echoes beyond that distance. In fact, some 600 km to the northeast of Taiwan, the same band brought a squall line with heavy precipitation and frequent lightning to Okinawa, starting around 0930 UTC. This storm prevented the P-3 from fueling and taking off until 1750 UTC, well after the SMMR pass and after the decay of the most significant precipitation within radar range of Hualien.

Several areas of echo that have little corresponding signature in the SMMR $\bar{P}$ field are noteworthy. They include several areas of weak convective showers south of the main band, plus a rather extensive area of very weak radar reflectivity between the 100- and 200-km range. The former discrepancy is likely to be due to the small size of the cells compared with the SMMR field of view (FOV).

The most plausible explanation for the latter is that the radar is picking up an area of widespread stratiform virga in the wake of the convective line. During P-3 flights through decaying precipitating systems, a relatively strong bright band was frequently observed at the melting level (~ 4.7 km above sea level) on the tail radar display, with only weak precipitation echoes extending a kilometer or less below the bright band. Not only was the effective depth of the precipitation layer very shallow so that the expected 37-GHz opacity due to rain would be much lower than for a rain layer extending to the surface, but the source cloud was usually observed to be almost completely glaciated so that cloud liquid water would also be expected to contribute negligibly to decreases in SMMR $\bar{P}$. Thus, the SMMR response to this cloud structure could be quite weak, while the radar, whose beam intercepts the bright band at these distances owing to the earth's curvature, could record disproportionately large reflectivities. In summary, we surmise that the actual area coverage by surface precipitation may be considerably lower than would be inferred from the radar echoes in this particular case, and that the vertically integrated attenuation inferred from the SMMR is likely to be the more reliable of the two indicators.

The 20 May case occurred between IOPs, while a weak Mei-Yu front was quasi-stationary over the north end of Taiwan. Soundings, however, were convectively unstable and very moist, and a large mesoscale convective system (MCS) developed about 200 km east of Taiwan, as seen in both the radar and the SMMR data (Fig. 6). Overall, there is excellent agreement between the SMMR and radar concerning both the shape and location of the major precipitation features. As before, some of the smaller, weaker showers seen by radar fail to show up in the SMMR data. Also, there are some areas beyond the 250-km range where the

¹ It has been found, when independent estimates of integrated water vapor and surface-wind roughening are available (e.g., from other channels of the microwave radiometer), that it is more efficient and reliable to use empirical relationships to estimate the clear-sky 37-GHz polarization difference directly from these parameters rather than resorting to the spatial statistical technique described by PK. Such an algorithm has been described for the newer SSM/I instrument by Petty (1990; see also Petty and Katsaros 1990b). Unfortunately, at the time of the TAMEX field phase, the principal SMMR water vapor channel was no longer functioning, and consequently, this approach could not be adapted to the SMMR data used here.

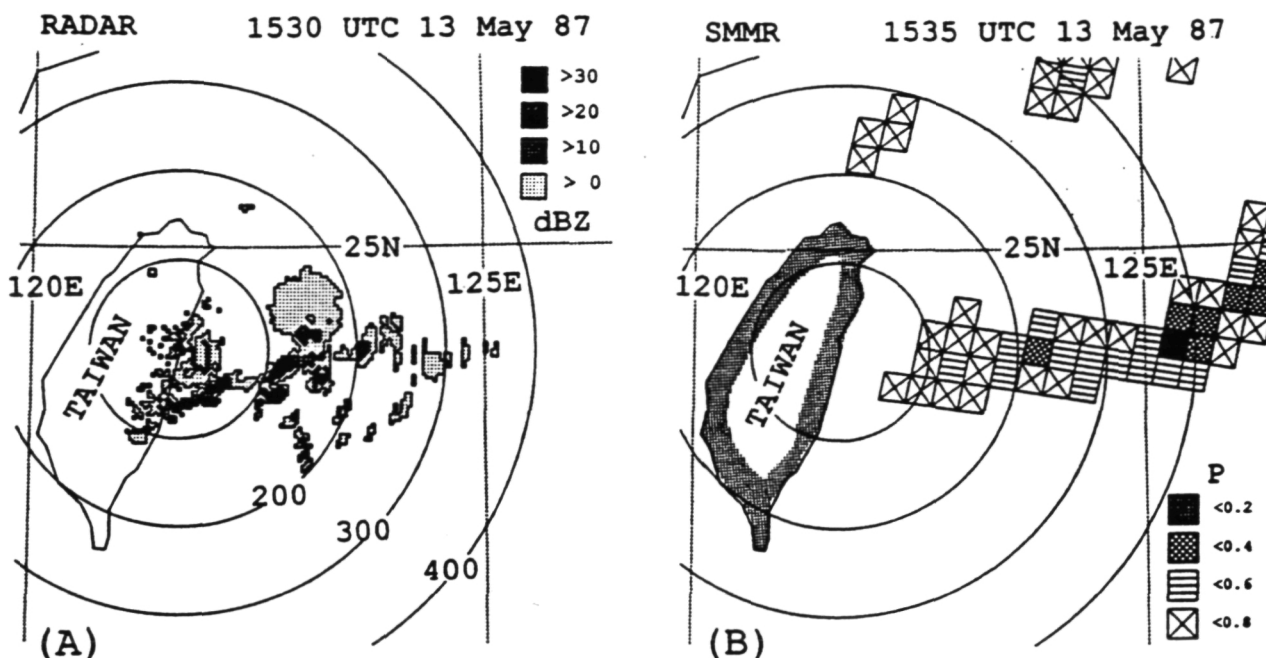


FIG. 5. (a) Hualien radar reflectivity (uncorrected) and (b) SMMR normalized 37-GHz polarization field for 13 May 1987 case. Times of observations are 1530 and 1535 UTC, respectively.

SMMR picks up precipitation that is apparently missed by the radar.

On 5 June (Fig. 7), an MCS developed on the east side of Taiwan and was responsible for the initiation of IOP 7 at 1200 UTC. The P-3 arrived in the area at approximately 1500 UTC and encountered extensive,

predominately stratiform precipitation near 22.5°N, 122.2°E. In anticipation of the SMMR overpass at 1539 UTC, the P-3 began flying a counterclockwise box pattern with roughly 50-km legs beginning at 300 m in altitude and ascending to 6100 m before descending to 300 m again and finally ending the pattern at

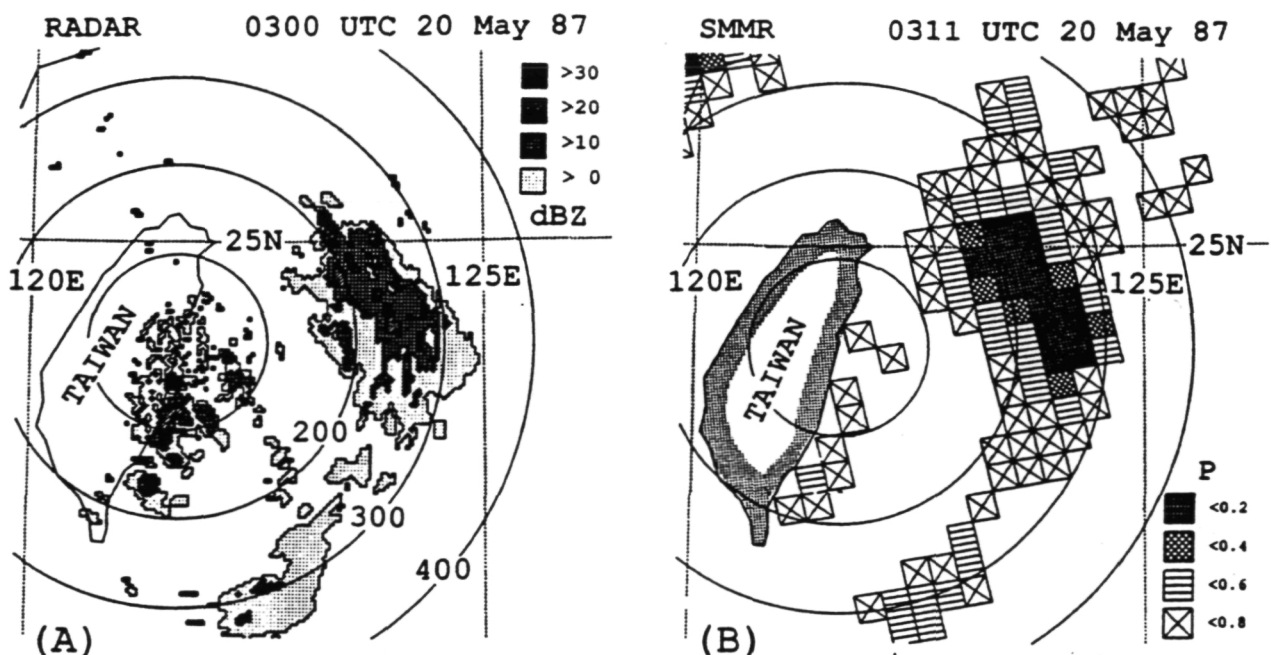


FIG. 6. (a) Hualien radar reflectivity (uncorrected) and (b) SMMR normalized 37-GHz polarization field for 20 May 1987 case. Times of observations are 0300 and 0311 UTC, respectively.

an altitude of 3000 m at 1557 UTC. The purpose was to obtain coincident vertical profiles of precipitation, cloud liquid water, and other microphysical parameters on a horizontal scale roughly comparable to the SMMR 37-GHz resolution. At the beginning of the pattern, rain in the vicinity of the P-3 was quite widespread and of moderate intensity, as inferred from the lower fuselage, tail, and nose radar returns. However, by the time of the SMMR overpass, little stratiform precipitation remained within 50 km of the aircraft. The tail radar, scanning vertically, continued to observe a distinct bright band at the melting level (4700 m) throughout much of the area, but only scattered echoes remained detectable below the bright band, most weaker than 20 dBZ, and many did not appear to extend all the way to the surface. In fact, both the Hualien radar and the SMMR indicate that a localized hole had formed over the sounding area within an otherwise extensive region of precipitation.

On 17 June (Fig. 8), a mesocyclone that had formed the previous night moved northward along the east coast. By evening, the system had begun moving eastward over the open ocean after bringing heavy precipitation ($>150 \text{ mm h}^{-1}$) to the northeast coast of Taiwan, apparently aided by considerable orographic enhancement. Over open ocean, Hualien radar reflectivities suggest more ordinary rain rates. At the time of the SMMR overpass at 1551 UTC (2351 LST), the P-3 was en route to the southwest end of the most active portion of the system (near 24°N , 123°E), which it subsequently penetrated at about 1610 UTC. This system proved to be one of the more vigorous and long-lived of those investigated by the P-3 during TAMEX. Its structure, orientation, and evolution in many ways resembled those of an intense midlatitude MCS. The Hualien surface radar, the P-3 lower fuselage radar, and the SMMR all agree on the general locations of the principal precipitation features; however, the surface radar and the P-3 disagree somewhat on the detailed structure, presumably because of the former's rather poor spatial resolution at that great range. It is not surprising, given the SMMR's coarse resolution, that it responded more strongly to the extensive stratiform precipitation in the older northeast portion of the system than to the narrow line of intense convective precipitation to the southwest.

6. Results

For each of the four cases described in section 5, the Hualien radar reflectivities, corrected according to (2)–(5), were converted to rain rate using the Z – R relationship given by (1). A roughly circular window encompassing 37 radar pixels was used to calculate FOV-averaged rain rate $\bar{R}$ and fractional coverage F_Z by echo exceeding various reflectivity thresholds for each SMMR pixel. In all, the comparison dataset consists of 518 radar–SMMR pairs, 264 of which contain at

least some radar echo. The results appear in Figs. 9–12. In each figure, the digits plotted indicate the distance of the corresponding radar–SMMR pair from the Hualien radar site (see figure legend).

In contrast to the results found by PK for two tropical cloud clusters over the South China Sea, there is a recognizable correlation in Fig. 9 between pixel-averaged rain rate $\bar{R}$ and $\bar{P}$, with higher values of $\bar{R}$ tending to be associated with lower values of $\bar{P}$ and vice versa. There is, however, considerable scatter in the relationship, but in view of the reasonably random distribution of range codes in Fig. 9, it does not appear that this scatter can be directly attributed to residual range-dependent errors in the radar rain rates. Rather, it is likely that much of the scatter arises from varying degrees of inhomogeneity in the rain-rate field within each FOV. As pointed out in numerous earlier studies (e.g., Wilhelm 1986; Chiu et al. 1990; Short and North 1990), nonlinearity in the relationship between rain rate and many microwave observables (including P) lead to ambiguities in rain-rate retrievals when rain is not uniformly distributed within the FOV. A theoretically derived rain-rate estimate based on the assumption of a uniformly filled FOV in fact serves only as a lower bound on the true FOV-averaged rain rate.

Nevertheless, the results in Fig. 9 appear to confirm that a statistically useful relationship exists between 37-GHz polarization observations and mean rain rate, at least for the conditions encountered in TAMEX. How to accurately quantify the relationship is less obvious, given the large amount of scatter and the relatively small number of pixels exhibiting larger rain rates. This problem will be addressed shortly.

In contrast to the results of PK, a simple straight-line relationship is not found between $\bar{P}$ and fractional coverage by echo at any reflectivity threshold (Figs. 10–12). However, we do find that a simple physical argument explains the approximate left and right boundaries on the clouds of points in these figures: If we postulate a unique monotonic relationship between reflectivity Z at a given radar grid point and local P , as seen by a hypothetical satellite sensor with the same spatial resolution as the radar, then the FOV-averaged polarization $\bar{P}$, seen by a real sensor with finite resolution (like the SMMR), is easily computed by averaging $P(Z)$ over the distribution of echo intensities within the FOV.

Under this assumption, if the fractional coverage of the FOV F_Z by radar reflectivities exceeding some threshold Z is known then the upper and lower bounds on $\bar{P}$ for that FOV are given by

$$\bar{P}_{\min} = (1 - F_Z)P(Z), \quad (6)$$

$$\bar{P}_{\max} = (1 - F_Z) + F_Z P(Z). \quad (7)$$

These expressions may be understood simply by considering the minimum or maximum polarizations that the fractions F_Z and $(1 - F_Z)$ of the FOV may have

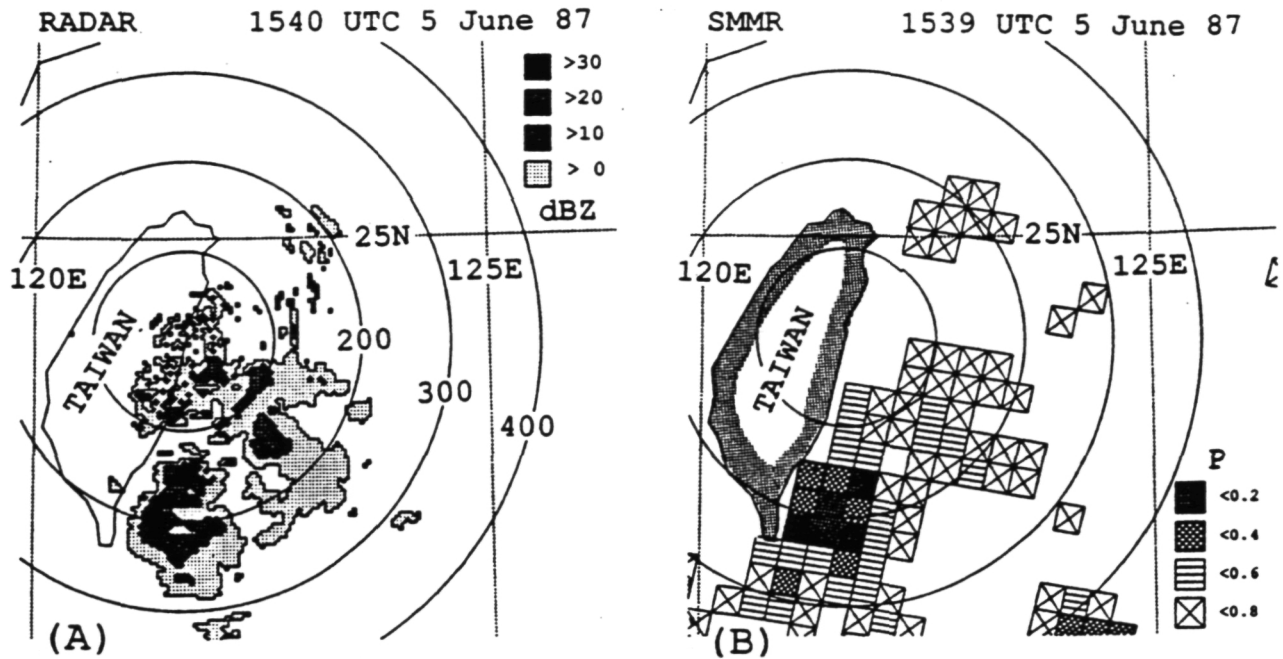


FIG. 7. (a) Hualien radar reflectivity (uncorrected) and (b) SMMR normalized 37-GHz polarization field for 5 June 1987 case. Times of observations are 1540 UTC and 1539 UTC, respectively.

in each case and then taking the weighted average for the entire FOV. For $\bar{P}_{\min}$, a fraction F_Z has a polarization of zero while the remainder has a polarization $P(Z)$: for $\bar{P}_{\max}$, the fraction F_Z has a polarization $P(Z)$ while the remainder has a polarization of unity.

Constraints of the form suggested by (6) and (7) appear to hold surprisingly well in each of Figs. 10–12, and from these plots alone one could easily make a plausible guess as to the approximate value of P corresponding to reflectivity thresholds of 0, 10, and 20

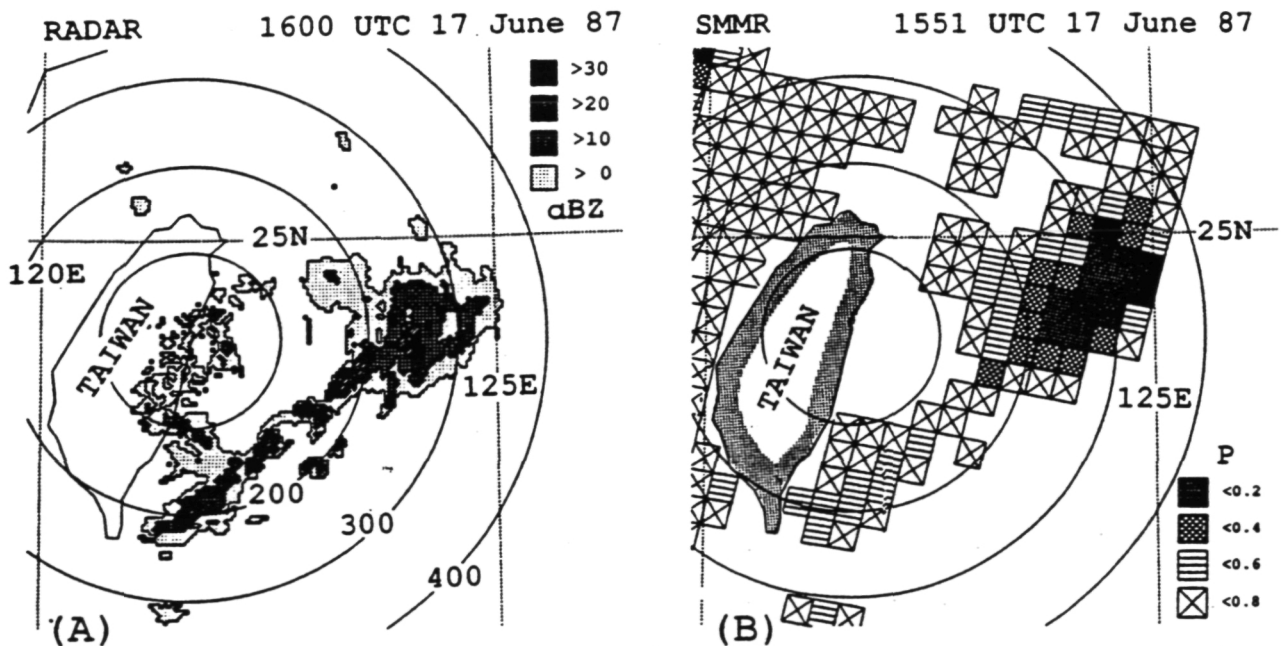


FIG. 8. (a) Hualien radar reflectivity (uncorrected) and (b) SMMR normalized 37-GHz polarization field for 17 June 1987 case. Times of observations are 1600 UTC and 1551 UTC, respectively.

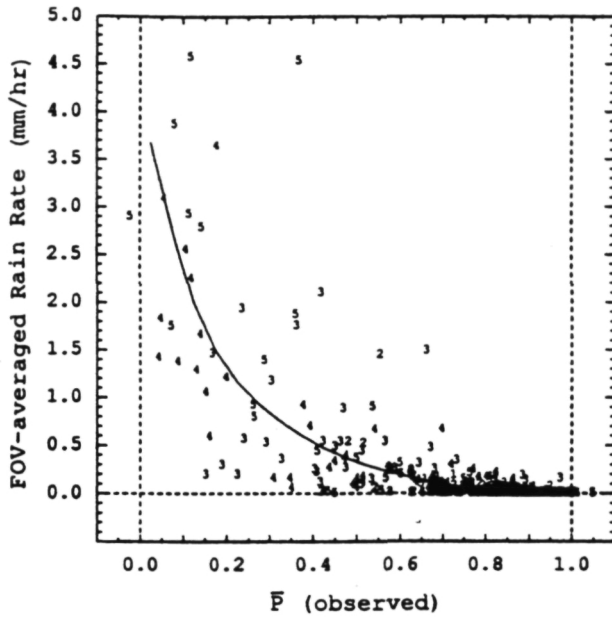


FIG. 9. Radar-estimated FOV-averaged rain rate $\bar{R}$ plotted as function of SMMR normalized 37-GHz polarization P . Digits plotted indicate the distance of each SMMR-radar pair from the Hualien surface radar site: 1 = 50–100 km, 2 = 100–150 km, 3 = 150–200 km, 4 = 200–250 km, 5 = 250–280 km. Solid curve represents mean FOV-averaged rain rate as a function of P , corresponding to column R_1 in Table 3.

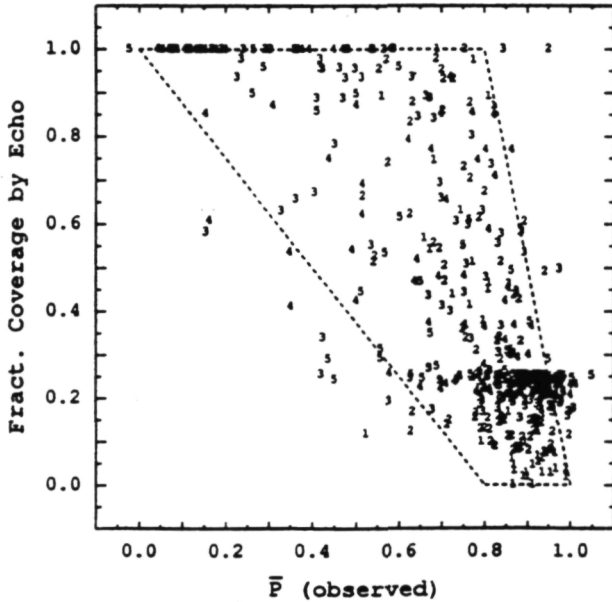


FIG. 10. Fractional coverage F_z of SMMR FOV by radar echo (recorded dBZ > 0), plotted as function of SMMR normalized 37-GHz polarization P . Digits plotted indicate range from Hualien as described for Fig. 9. Dashed diagonal lines indicate the theoretical upper and lower bounds on P indicated by (6) and (7). Radar data were corrected using (2)–(5) to account for echo area loss at greater ranges.

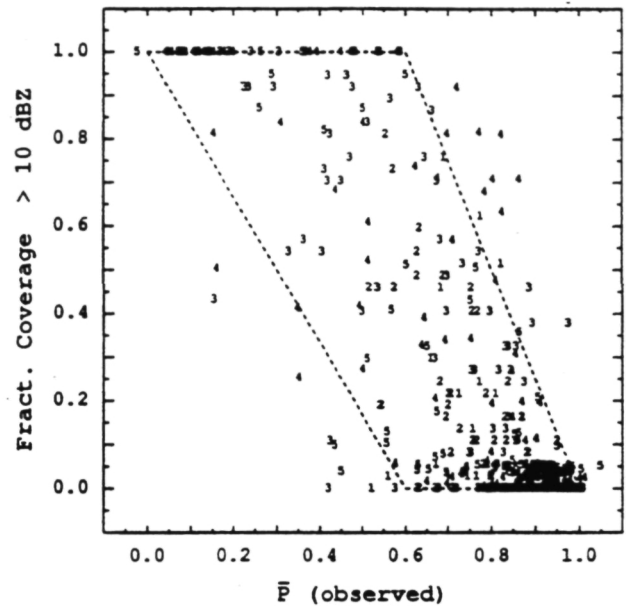


FIG. 11. Same as Fig. 10, but for a reflectivity threshold (corrected) of 10 dBZ.

dBZ. This result led to the search for a P – Z relationship that would allow the SMMR-observed $\bar{P}$ to be simulated from the radar reflectivity fields.

As pointed out by PK, for a locally plane-parallel atmosphere bounded by a nearly specular sea surface, P is well-approximated by the square of the slant-path rain-cloud transmittance τ , or

$$P \approx \tau^2 = \exp \left[\frac{-2\sigma}{\cos(\theta)} \right], \quad (8)$$

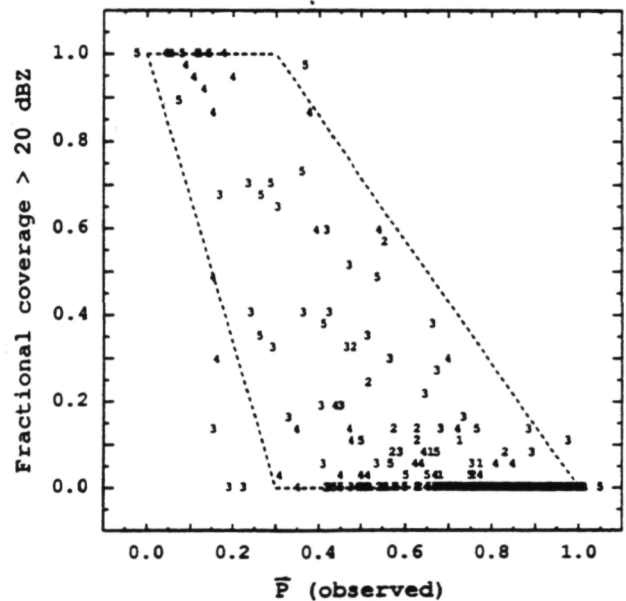


FIG. 12. Same as Fig. 10, but for a reflectivity threshold (corrected) of 20 dBZ.

where σ is the vertically integrated microwave opacity contributed by liquid hydrometeors, and $\theta \approx 50^\circ$ is the viewing angle of the SMMR. This relationship was first inferred from a simple algebraic treatment of the atmospheric and surface components of the polarized radiation seen from space. Recent numerical experiments using the more accurate microwave radiative transfer model of Petty (1990) confirm that (8) holds to better than 5%, except for small values of τ .

The total liquid water optical depth σ may be approximated as the product of the rain-layer depth H and the mean volume extinction coefficient $k_{e,R}$ corresponding to the rain rate R . In addition, there may be a significant contribution to σ from the column-integrated nonprecipitating cloud-water content L . We may therefore write

$$\sigma = k_{e,R}H + \kappa_{e,L}L, \quad (9)$$

where $\kappa_{e,L}$ is the average mass extinction coefficient of nonprecipitating cloud liquid water.

Savage's (1976) Mie extinction calculations for a Marshall-Palmer distribution of drop sizes yield a power-law relationship between $k_{e,R}$ at 37 GHz and R ,

$$k_{e,R} \text{ (km}^{-1}\text{)} = 0.077R^{0.97}, \quad (10)$$

assuming a rainwater temperature of 10°C . By further combining this with the Marshall-Palmer Z - R relationship,

$$Z = 200R^{1.6}, \quad (11)$$

a direct relationship between 37-GHz extinction and radar reflectivity factor is obtained:

$$k_{e,R} \text{ (km}^{-1}\text{)} = 3.10 \times 10^{-3}Z^{0.606}. \quad (12)$$

This relationship is strictly valid only for a Marshall-Palmer size distribution, but is likely to be a useful approximation for most other realistic size distributions as well.

The calculations of Petty (1990) indicate that the 37-GHz mass-extinction coefficient $\kappa_{e,L}$ of cloud liquid water is $0.267 \text{ m}^2 \text{ kg}^{-1}$, assuming that the temperature of the bulk of the liquid water is near 0°C . All that remains is to specify how L varies with Z (or, equivalently, with R). Such a relationship cannot be derived from first principles, nor is there much observational data to provide guidance on this matter. Therefore, a linear relationship between L and $\log Z$ was chosen subjectively, and the coefficients that maximized the agreement between the P - Z model and the SMMR-radar comparison data were then determined:

$$L \text{ (kg m}^{-2}\text{)} = 0.2 + 0.02\text{dBZ}. \quad (13)$$

According to the above model, 0.2 kg m^{-2} is the average cloud liquid water present for the minimum recorded radar reflectivity $\text{dBZ} = 0$. For a relatively intense echo of 40 dBZ ($R \approx 15 \text{ mm h}^{-1}$), the assumed cloud liquid water content takes on a value of 1.0 kg m^{-2} .

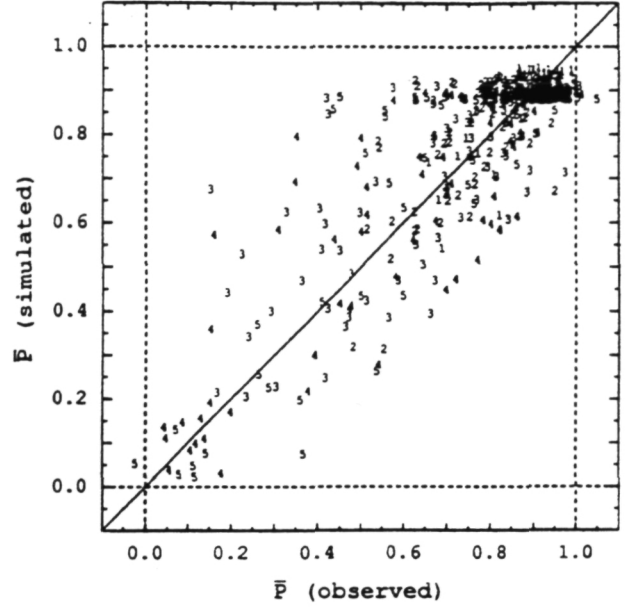


FIG. 13. Simulated $\bar{P}$, computed from radar data using (14) and averaged over SMMR footprint, plotted as function of SMMR-observed $\bar{P}$. Range codes are as described for Fig. 9.

Combining (8), (9), (12), and (13) and assuming $\theta = 50^\circ$ and $H = 4.5 \text{ km}$, we obtain the following closed expression for P as a function of Z :

$$P = 0.847Z^{-0.0722} \exp(-0.0434Z^{0.606}). \quad (14)$$

Applying (14) to reflectivity thresholds of 0, 10, and 20 dBZ and invoking (6) and (7) yields the theoretical upper and lower bounds on $\bar{P}$ indicated by the dashed lines in Figs. 10–12. In each plot, only a very small fraction of the 518 total points is seen to fall significantly outside these boundaries.

A theoretical value of $\bar{P}$ for each SMMR observation was calculated by computing P from the radar data using (14) and averaging over the field of view. To reduce biases arising from variable cloud water at non-raining radar grid points, P was allowed to vary randomly at those points between 0.8 and 1. The simulated $\bar{P}$ values are plotted against the observed $\bar{P}$ in Fig. 13. The general agreement is quite satisfactory, apart from a moderate degree of scatter that may or may not actually have much to do with the validity of this approach for simulating $\bar{P}$. Possible sources and implications of the random differences between theory and observation will be considered in more detail in the remainder of this section.

Given that (14) appears to be a reasonable model for the relationship between P and Z , we are no longer limited solely to the coincident radar-SMMR data for deriving a statistically unbiased empirical relationship between FOV-averaged rain rate $\bar{R}$ and SMMR-observed $\bar{P}$. Indeed, using a larger set of radar scans, it is

now possible to produce a simulated radar-SMMR dataset that is fully consistent with the direct comparisons but that also permits the statistical variability induced by spatially inhomogeneous rainfall within SMMR FOVs to be examined in much greater detail. This is in essence a Monte Carlo approach to examining the mean relationship between $\bar{R}$ and $\bar{P}$, but uses real radar data, rather than a stochastic model, to generate the required rainfall spatial statistics.

By calculating simulated $\bar{P}$ from all 390 Hualien radar scans, two-dimensional histograms of simulated $\bar{P}$ versus radar estimated $\bar{R}$ were generated. The same SMMR FOV shape and area were used, as in the direct comparisons, but the FOV was allowed to be centered, on every radar grid point so that the number of $\bar{P}$ - $\bar{R}$ pairs approximately equaled the total number of radar pixels used. In order for the spatial characteristics of the echoes to reflect the actual structure of the precipitation more accurately, only radar data between the 50- and 150-km range from Hualien were used so that the effects of surface clutter, antenna beamwidth, and echo truncation are reduced. The resulting sample consisted of 5.8×10^5 pairs, of which 3.1×10^5 of the simulated FOVs contained precipitation.

Histograms and related statistics were calculated twice, each time under opposing assumptions about the nature of the random differences between predicted and observed $\bar{P}$ evident in Fig. 13. For the first run, it was assumed that (14) correctly predicts local P as a function of local Z and that all of the scatter in Fig.

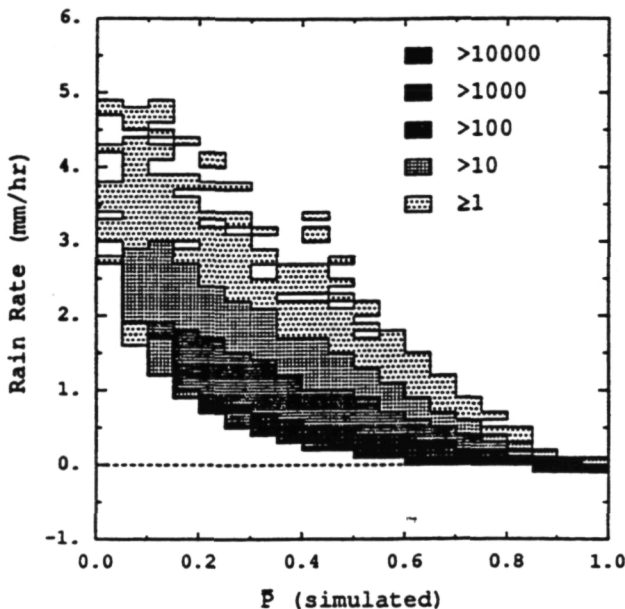


FIG. 14. Bivariate histogram of simulated $\bar{P}$ and FOV-averaged rain rate $\bar{R}$, computed from all Hualien radar scans archived during TAMEX period. Only data from 50 to 150 km from Hualien were used. Equation (14) is assumed here to be exact. Resolution of the histograms is 0.1 mm h^{-1} in rain rate and 0.05 in P . Shading indicates the number of occurrences recorded for each histogram bin.

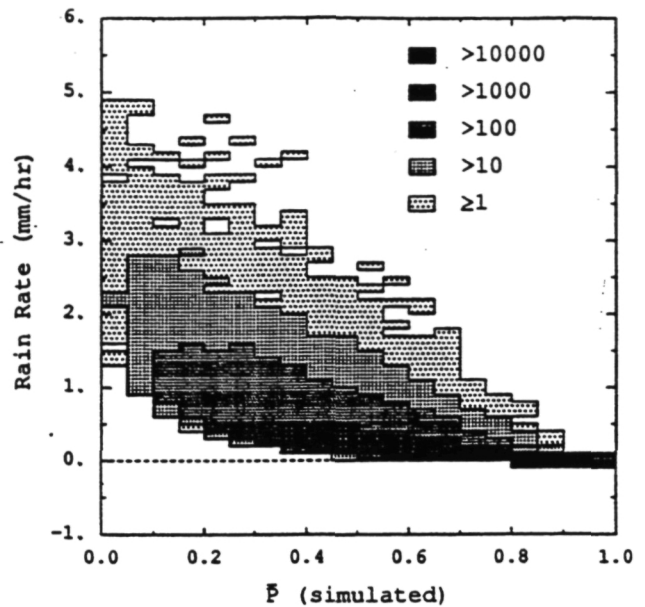


FIG. 15. Same as Fig. 14, but the simulated $\bar{P}$ include a component of random uncertainty (see text).

13 reflects only the inevitable errors in the collocation of SMMR and radar fields and/or the modest time differences between the two datasets. For the second run, it was assumed that most of the scatter may be attributed to intrinsic statistical variability in the relationship between reflectivity patterns within each SMMR FOV and the observed $\bar{P}$ due to variability in cloud geometry, nonprecipitating liquid water contribution, drop-size distribution, or some other physical factor. Also counted in this category are quasi-random errors in $\bar{P}$ arising from the simple algorithms used to process and grid the SMMR brightness temperatures (R. Spencer, personal communication 1989).

Thus, $\bar{P}$ - $\bar{R}$ pairs were simulated from the radar data for the second run exactly as for the first, but the $\bar{P}$ results were then modified by a random noise factor that approximately mimics the scatter in Fig. 13. This was accomplished by raising the exact $\bar{P}$ to the power e^u , where u is a uniform random deviate lying in the range ± 0.405 . This approach is equivalent to assuming a factor of $(1.5)^{\pm 1}$ uncertainty in the FOV effective rain-cloud optical depth. Values of $\bar{P}$ near zero or near unity are relatively unaffected by such uncertainties, whereas intermediate values may be substantially modified.

The resulting histograms for both sets of calculations are shown in Figs. 14 and 15. Both are characterized by a sharply defined minimum $\bar{R}$ for a given value of $\bar{P}$. The lower limit in the first corresponds to the expected result for uniform rain rate within the FOV and is determined exactly by the assumed P - Z relation (14) and Z - R relation (1). In the second histogram, the lower limit is modified by the assumed uncertainty

TABLE 3. Computed mean FOV-averaged rain rate (mm h^{-1}) corresponding to intervals of 37-GHz P . Column R_1 , rain rate computed from deterministic P simulations; column R_2 , rain rate computed from simulations including random uncertainties in P .

P	R_1	R_2
0.00–0.05	3.68	2.76
0.05–0.10	2.75	1.87
0.10–0.15	1.99	1.41
0.15–0.20	1.50	1.18
0.20–0.25	1.16	0.99
0.25–0.30	0.93	0.83
0.30–0.35	0.75	0.69
0.35–0.40	0.60	0.57
0.40–0.45	0.47	0.46
0.45–0.50	0.37	0.37
0.50–0.55	0.29	0.29
0.55–0.60	0.22	0.23
0.60–0.65	0.17	0.17
0.65–0.70	0.12	0.13
0.70–0.75	0.08	0.09
0.75–0.80	0.05	0.05
0.80–0.85	0.03	0.02
0.85–0.90	0.00	0.00
0.90–0.95	0.00	0.00
0.95–1.00	0.00	0.00

in P for a given value of R . In this case, substituting $P^{2/3}$ for P in (14) determines $\bar{R}_{\min}(\bar{P})$.

Ensemble average rain rates were computed for each interval of $\bar{P}$ and are tabulated in Table 3. Since the two sets of computations represent limiting extremes with regard to the assumed nature of the scatter appearing in Fig. 13, the values in Table 3 give a good indication of the range of uncertainty in the mean rain rate corresponding to a given value of $\bar{P}$. For larger values of $\bar{P}$, the difference between the two approaches is very small and probably of no practical importance. The greatest fractional uncertainty is for $\bar{P} < 0.3$, for which the maximum error is still generally below about 25%.

To show that the results from the simulation are consistent with the original direct comparisons between observed $\bar{P}$ and FOV-averaged rain rate, a curve corresponding to column R_1 in Table 3 is superimposed on Fig. 9. Despite the rather indirect approach used in deriving this $\bar{P}$ – $\bar{R}$ relationship, it would clearly be difficult to justify a significantly different curve fit to the SMMR–radar comparison data.

Finally, to give an indication of how the larger uncertainties in $\bar{R}$ for low values of $\bar{P}$ might affect mean rain rates retrieved over a large area or long time period, the fraction of total rainfall accounted for $\bar{P}$ between 0.0 and an arbitrary upper threshold has been plotted (Fig. 16). Less than 30% of the total rainfall in this dataset is associated with the small values of $\bar{P}$ for which the distinction between the two columns in Table 3 may be important.

It bears reemphasizing that the aforementioned uncertainties are those associated with the ensemble av-

erage rain rate associated with a given value of $\bar{P}$; that is, they reflect the potential for biased average rain rates arising from incorrect assumptions about the nature of the residual scatter in Fig. 13. While these potential biases appear to be quite modest, the histograms in Figs. 14 and 15 clearly show that, on a pixel-by-pixel basis, estimates of $\bar{R}$ based on 37-GHz polarization observations will have much larger uncertainties, often exceeding a factor of 2.

Referring to Fig. 16, it can also be seen that no more than about a 4% error in total retrieved rainfall will be made by regarding SMMR pixels with $\bar{P} > 0.8$ as completely precipitation-free; conversely, Fig. 9 suggests that only a small minority of pixels with $\bar{P} < 0.8$ are nonraining. Consequently, a threshold of $\bar{P} = 0.8$ would appear to offer a convenient and reasonably reliable basis for distinguishing between raining and nonraining pixels.

7. Discussion and conclusions

This study represents a continuation of work begun by Petty and Katsaros (1990a) to refine an attenuation-based technique for estimating rainfall parameters from polarized 37-GHz brightness temperatures. Because of the highly variable microphysical and mesoscale structure of precipitation, empirical comparisons between the microwave data and surface digital radar in the tropics and subtropics are a particularly important component of this effort, despite the relatively straightforward physical interpretation of our normalized polarization difference P .

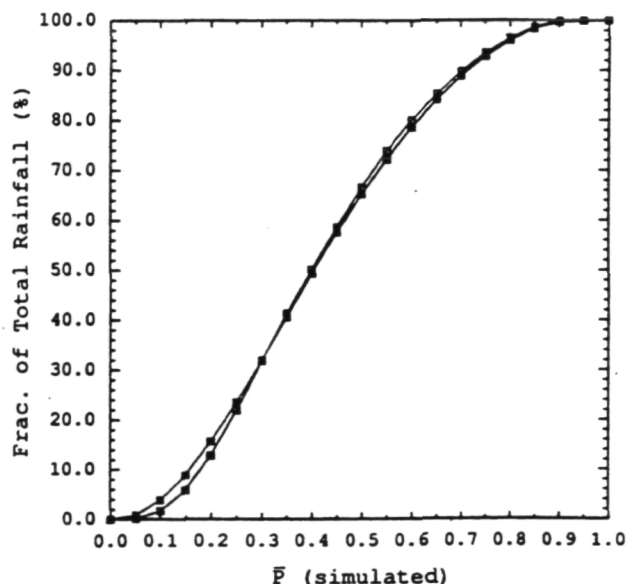


FIG. 16. Fraction of total area-integrated rainfall associated with model-predicted $\bar{P}$ less than or equal to indicated value. Open squares correspond to data appearing in Fig. 14; filled squares correspond to Fig. 15.

The comparison dataset obtained from the Hualien radar is considerably larger than the dataset from the Bintulu radar employed in PK because of the larger over-ocean area covered, the longer time period for which data were collected during TAMEX, and the fewer problems with surface obstructions. Most importantly, the TAMEX radar dataset was well suited to a thorough statistical evaluation of range-dependent errors, which allowed us to try to correct these prior to performing the comparisons with the SMMR observations of precipitation. To our knowledge, the empirical radar correction technique developed herein is novel, and it remains to be seen whether this technique can be applied with equal success to other over-water digital radar datasets.

Because of both the larger sample size and our more comprehensive examination of radar rain-rate errors, we have somewhat greater confidence in the results presented here as compared with those from our earlier radar-SMMR study, to the extent that the conclusions reached in the two studies disagree. For example, in the present study, we found that the relationship between the SMMR 37-GHz normalized polarization difference $\bar{P}$ and fractional coverage by precipitation was less deterministic than that observed by PK but that simple physical arguments succeed remarkably well in predicting effective constraints on this relationship. Also in contrast to PK, the TAMEX dataset allows a clear correlation between $\bar{P}$ and average rain rate to be discerned. Not unexpectedly, though, the uncertainty in $\bar{R}$ inferred from a single measurement of $\bar{P}$ remains rather large.

Apart from differences in the quality and size of the comparison datasets used by PK and in the present study, significant differences in the physical characteristics of the precipitating systems themselves may be responsible for some differences in our results. For example, the characteristic rain rates observed by the Hualien radar during TAMEX were rather low, as indicated by the cumulative reflectivity distribution [Fig. 4 and Eq. (3)]. At the spatial resolution of the radar data, 50% of the total rainfall was contributed by rain rates less than 1.4 mm h^{-1} , and 75% was contributed by rates less than 3.5 mm h^{-1} . Although comparable statistics were not calculated for the equatorial convection studied by PK, somewhat higher maximum echo intensities were observed in the WMONEX cases.

Furthermore, it was our observation, based both on the radar data and on our participation in the P-3 flights, that a relatively high proportion of the total rainfall was associated with the stratiform remnants of prior convection. With the exception of a few organized, moderately intense propagating squall lines, convection was often observed to decay rather rapidly after reaching its peak, leaving large, quasi-stationary masses of stratiform precipitation in its place. The relative predominance of stratiform precipitation, as compared with the PK data, would explain not only

the low average rain rates but also the stronger correlation of $\bar{P}$ with rain rate observed in this study, since saturation of the 37-GHz signal would be expected to occur for a smaller proportion of the total rainfall.

In order to arrive at a reasonably confident specification of FOV-averaged rain rate $\bar{R}$ as a function of observed $\bar{P}$, we found it necessary to greatly increase the effective sample size. This was accomplished by simulating $\bar{P}$ from the large set of TAMEX radar scans for which there were no coincident SMMR observations. The P - Z relation used for this purpose was derived theoretically and was not tuned except for the empirical specification of nonprecipitating cloud water as a function of Z . Despite the simplicity of the model and the fact that no comprehensive radiative transfer calculations were involved, we found that $\bar{P}$ - $\bar{R}$ statistics derived from the simulated dataset were remarkably consistent with the direct observations. The problem of variable footprint filling on microwave rain-rate retrievals was examined quantitatively via this method, and the results are summarized by the histograms in Figs. 14 and 15 and by Table 3.

In the absence of geophysical noise other than that arising from variable footprint filling, the ensemble averaged $\bar{R}$ values computed for intervals of observed $\bar{P}$ (Table 3, column labeled R_1) are approximately 30% greater than the values that would have been retrieved from (11) and (14) under the assumption of horizontally homogenous rain. This difference is about equal to the beam-filling bias predicted theoretically by Chiu et al. (1990), but is significantly less than the factor-of-2 bias determined empirically by Short and North (1990) for Electrically Scanned Microwave Radiometer (ESMR) observations of GATE-type rainfall. However, given the significant differences in the frequency used (37 GHz vs 19.35 GHz) and the variable used for rain-rate retrieval (normalized polarization vs brightness temperature), there is no reason to expect more than qualitative similarity between the results of the above studies and those presented here.

At least for meteorological conditions similar to those encountered during TAMEX, the $\bar{P}$ - $\bar{R}$ relationship tabulated in Table 3 should permit reasonably unbiased estimates of FOV-averaged rain rate using SMMR 37-GHz P values derived from gridded (Cell-All) SMMR brightness temperatures. The relationship incorporates both the physical dependence of P on rain rate at a given point and the spatial statistics of TAMEX rainfall on scales smaller than the resolution of the SMMR 37-GHz channels. Where the spatial characteristics differ substantially, these results may lead to biased rainfall totals.

As already noted, the freezing level, and hence the approximate depth of stratiform rain, was near 4.5 km throughout TAMEX. This value is typical of much of the tropics and subtropics; hence, little or no adjustment should be necessary for freezing level when applying the TAMEX results to those regions. In cooler

regimes, lower freezing level implies lower stratiform rain-cloud opacity (and hence higher P) for a given rain rate. Under these conditions, some adjustment will undoubtedly be necessary, though the greater spatial homogeneity of stratiform rain may partially compensate for this effect.

One important remaining source of uncertainty that must be acknowledged lies in the Z - R relationship (1) employed to estimate rain rates from the radar data. It is well known that the actual relationship between radar reflectivity and rain rate can vary rather dramatically from location to location and from storm to storm (Joss and Waldvogel 1990). Indeed, a great number of Z - R relationships have appeared in the literature, each based either on empirical comparisons between radar data and rain gauges under differing meteorological conditions or on theoretical calculations entailing differing assumptions about drop-size distributions. Overall, we estimate that up to a factor-of-2 uncertainty might be introduced in any given case by our lack of knowledge of the best Z - R relationship.

Since July 1987, the newer SSM/I has been operational aboard Defense Department satellites. Like the SMMR, this new microwave imager includes dual-polarization 37-GHz channels, with only minor differences from the SMMR with respect to nadir angle and surface resolution. With slight modification, the general results obtained here for the SMMR 37-GHz channels should therefore be applicable to the SSM/I as well. We recommend, however, that SSM/I precipitation estimates eventually be calibrated directly against surface radar, as was done here for SMMR, and that these comparisons be done not only for the 37-GHz channels but for the 19.35- and 85.5-GHz channels of this instrument as well. We expect that improvements in the calibration, sampling characteristics, and processing of the SSM/I data will permit much more reliable and useful precipitation algorithms to be developed for the SSM/I than would have been possible for the SMMR. Ultimately, a suitable combination of the information obtained from normalized 19-, 37-, and 85-GHz polarization differences, plus the 85-GHz volume-scattering signal described by Spencer et al. (1989) and refined by Petty (1990), may offer the best hope for overcoming the limitations in dynamic range and in sensitivity and/or spatial resolution associated with each SSM/I frequency individually.

The Hualien site remains part of the operational radar network belonging to the Republic of China's Central Weather Bureau and should be considered a good candidate for future empirical studies of microwave signature of precipitation, using the SSM/I and its successors. The site's attractiveness is increased by the fact that precipitation observed in that region may take on either a tropical or midlatitude character, depending on season and weather regime, so that empirical relationships derived there may have broader applicability than might be expected elsewhere.

Acknowledgments. We would like to thank the TAMEX project leaders, Drs. Ying-Hwa Kuo and George Tai-Jen Chen, for their gracious accommodation of our "piggy-back" involvement in the TAMEX field phase. Thanks also to Dr. David Jorgensen of NOAA for his cooperation and assistance in every aspect of our participation. Dr. Dean Churchill made helpful suggestions regarding the evaluation and correction of the radar data. Hualien radar data were provided by the U.S. TAMEX Project Data Management Office at NCAR. *Nimbus-7* SMMR Cell-All data were processed by the SMMR Information Processing Team of the Nimbus Project and provided to us by the National Space Science Data Center at Goddard Space Flight Center. This work was initially funded by NOAA Grant 40AANE701626, administered by Dr. John Alishouse, whose genuine interest is much appreciated. Additional support has come from NASA Grant NAG5-943.

APPENDIX

Aircraft Liquid Water Compared with SMMR P

At the time of the SMMR pass for the 17 June case described in section 5, the P-3 was executing a straight-line ascent through a stratiform cloud layer with bases below 300 m and tops at 2000 m. The horizontal distance traversed during the ascent through the cloud layer was 25 km, or approximately the diameter of a single SMMR 37-GHz pixel. The Johnson-Williams liquid water probe measured an average 0.32 g m^{-3} within the cloud layer, implying a column integrated liquid water content of 0.54 kg m^{-2} . No further cloud water was observed between 2000 m and the P-3's final altitude of 3170 m reached before penetrating the MCS a short time later.

The opportunity to compare even crude in situ measurements of integrated cloud liquid water L over the ocean with simultaneous satellite microwave observations is exceedingly rare. Therefore, the relationship

$$L (\text{kg m}^{-2}) = -2.06 \log_e(P) \quad (\text{A1})$$

is used to estimate integrated liquid water amounts corresponding to the SMMR P values observed in the immediate vicinity of the P-3, and these are compared with the P-3 estimate. The relationship (A1) was derived, as described by Petty (1990; see also Petty and Katsaros 1990b), for SSM/I 37-GHz retrievals of liquid water in marine stratocumulus; however, the leading coefficient has been adjusted here by a factor of 1.45, mainly to account for the significant decrease of microwave absorption by cloud water that accompanies an increase of cloud temperature from near 285 to near 300 K, as well as to account for a 3° difference in viewing angle.

The P-3 began its ascent directly under a SMMR pixel for which an integrated liquid water value of 0.33

kg m^{-2} , based on $\bar{P} = 0.85$ was obtained. By the time the aircraft had ascended out of the cloud layer, it was approaching the center of the next SMMR pixel with estimated liquid water of 0.71 kg m^{-2} , based on $\bar{P} = 0.71$. The latter estimate assumes that no rain was present within the pixel, though some brightband return was evident on the tail radar display at that time.

Given the obviously strong horizontal gradients present in the integrated liquid water field, together with the coarse resolution of the SMMR, it is difficult to be certain that comparisons between flight-track measurements and the SMMR values are meaningful. Nevertheless, we note that the P-3 value of 0.54 kg m^{-2} is quite consistent with the average liquid water (based on equal weights) of 0.52 kg m^{-2} estimated from the two SMMR pixels that overlapped the 25-km track segment.

It must be acknowledged that such good agreement may very well be fortuitous, as the inherent uncertainties in the aircraft and satellite liquid water estimates probably exceed $\sim 10\%$ each, quite apart from the sampling incompatibilities between the two estimates. If anything, the fact that a six-week field experiment and ten aircraft missions, four of which coincided with SMMR overpasses, yielded only this one questionable comparison datum suggests that coincident aircraft liquid water and microphysical measurements offer an uneconomical and unreliable basis for developing and validating satellite liquid water and precipitation retrieval algorithms over the ocean. The prospects are likely to improve in a field experiment dedicated exclusively to algorithm validation, but the logistical problems and sampling difficulties remain immense.

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NEW GEOPHYSICAL ALGORITHMS FOR THE SPECIAL SENSOR MICROWAVE IMAGER

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1. Introduction

Although three years have now passed since the launch of the first Special Sensor Microwave Imager (SSM/I), there continues to be surprisingly few well-tested, well-documented, globally applicable algorithms for the estimation of atmospheric parameters over the ocean. Because of our interest in applying SSM/I data to case studies of weather systems over the ocean, we therefore undertook to assemble a unified set of algorithms which will permit us to extract the maximum amount of useful environmental information from raw SSM/I images, with particular attention to both the problems of sensor resolution and of cross-parameter contamination. While this aim remains only partially realized, particularly with regard to the estimation of rain parameters, the algorithms described below have already proven useful in our own work and are therefore perhaps of potential interest to others as well.

2. Background

The first operational algorithms — the so-called Hughes algorithms developed prior to the SSM/I's deployment in June 1987 — failed, in some cases spectacularly, to fulfill the performance criteria specified for each geophysical parameter, as was subsequently documented by the SSM/I Calibration/Validation Team (NRL, 1988). As part of the Cal/Val effort, new algorithms meeting specifications were derived for certain parameters, mainly using statistical regression of SSM/I brightness temperatures against *in situ* observations of the desired parameter. Because these new algorithms were primarily intended to replace the original Hughes algorithms within the framework of existing operational retrieval software of the Naval Oceanography Command and the Air Force Weather Service, they were required to meet the same severe computational constraints as were the Hughes algorithms; namely, that each algorithm should entail only linear combinations of at most four SSM/I channels.

Despite these constraints, and despite certain fundamental drawbacks inherent in a purely statistical approach to parameter retrieval, some of the replacement algorithms appear to perform satisfactorily for most applications. For example, the wind speed algorithm of Goodberlet et al. (1989), derived from simultaneous buoy wind reports representing a wide geographic area, appears to yield accuracies of about $\pm 2 \text{ m s}^{-1}$ under most conditions, provided that no significant precipitation is present within the field of view (FOV) of the lowest resolution channels used.

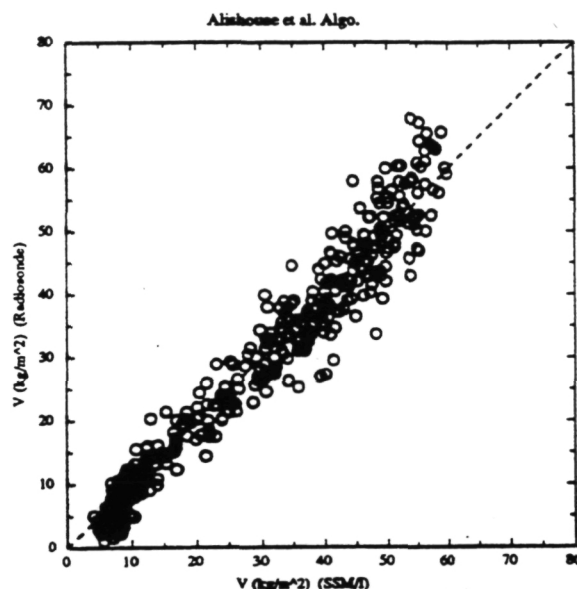


Fig. 1 SSM/I integrated water vapor V , obtained using the Alishouse et al. (1990) algorithm, plotted against nearly coincident radiosonde estimates of V . 515 data pairs were used, representing conditions ranging from tropical to polar.

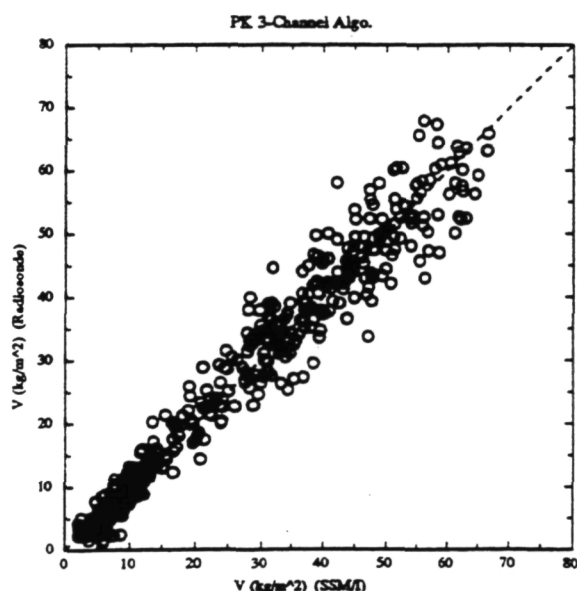


Fig. 2 Same as Fig. 1, but SSM/I retrievals are based on the water vapor algorithm (Eq. 2) presented in this paper. Note the absence of systematic biases at high and low values of V .

Alishouse et al. (1990) also successfully used statistical regression (SSM/I brightness temperatures vs. coincident radiosonde measurements of humidity) to obtain a global algorithm for integrated water vapor which met performance specifications, albeit only after departing from a purely linear algorithm formulation. However, although the performance of their algorithm is very good under most conditions, we have found that it exhibits systematic biases at very high and very low water vapor values, i.e., above $\sim 55 \text{ kg m}^{-2}$ and below $\sim 10 \text{ kg m}^{-2}$ (Fig. 1). Users of the Alishouse et al. (1990) algorithm are also cautioned by those authors to screen data carefully for precipitation, since they perceived an increased potential for contamination as a result of the nonlinear input term.

To the best of our knowledge, no completely satisfactory algorithm for SSM/I retrievals of cloud liquid water over the ocean has yet been demonstrated. As part of the Cal/Val effort, attempts have been made to generate new cloud water algorithms using the same statistical procedure as for integrated water vapor. In contrast to water vapor, however, coincident "surface truth" measurements of integrated cloud liquid water are extremely scarce, especially over the ocean. Cal/Val team members were thus able to obtain only a small set of surface-based observations of liquid water that could reasonably be compared with SSM/I data. These were made by upward-looking microwave radiometer at two locations: San Nicolas Island off the southern coast of California and Kwajalein Island in the tropics.

We tested the resulting preliminary four-channel and six-channel algorithms (Dr. John Alishouse, 1989; personal communication) by applying them to a large ensemble of SSM/I brightness temperatures which were determined from GOES visible imagery to represent essentially cloud-free conditions, so that the true liquid water content was known to be near zero. Unfortunately, we found that retrieved values were distributed over a range of over 30 kg m^{-2} and were highly correlated with retrieved integrated water vapor. We have concluded that the surface-based observations used to derive those algorithms represented too small a subset of possible meteorological conditions, and that a strong natural correlation between cloud water and integrated water vapor apparently led to regression relations which tend to equate high water vapor with high cloud water, even when no cloud water is in fact present.

Given the probable difficulty and expense of assembling a "surface truth" liquid water data set which would circumvent these problems, it seems unlikely that a successful liquid water algorithm can be achieved based solely on statistical regression of brightness temperatures against coincident measurements. The only alternative, and consequently the one we pursue in this paper, is to base retrievals at least in part on physical models of microwave radiative transfer.

Although the development of algorithms for rain parameter retrieval remains one of the most active areas of research within the microwave sensing community, accurate operational SSM/I retrievals of rain rate over the ocean have also not, to our knowledge, been demonstrated. Nevertheless, considerable progress has been made recently by a number of workers, both toward understanding the physics

of the interaction between rain clouds and microwave radiation and in obtaining empirical data on the microwave signature of rain. Now even some reasonably simple algorithms (e.g., Petty and Katsaros, 1990b; Spencer et al., 1989) appear to provide useful qualitative indices of the presence and intensity of precipitation, though absolute calibration of these algorithms remains a formidable task. Fortunately, for many applications, simple knowledge of the location and character of precipitation already represents a significant improvement in our ability to diagnose oceanic weather systems.

In this paper we present a self-contained set of algorithms for observing atmospheric water parameters and surface wind speed in weather systems over the ocean. Parameters retrieved include integrated water vapor, integrated cloud liquid water, surface wind speed, and the location and relative intensity of precipitation. For some of these parameters we adapt algorithms or techniques developed by other authors; our contribution in such cases is to incorporate those algorithms, with some minor modifications or enhancements, as components in a unified framework.

Note that in addition to the statistical algorithms developed within the context of the SSM/I Cal/Val project, physically-based retrievals of environmental parameters over the ocean are being distributed by Remote Sensing Systems, Inc. (Wentz, 1988). These are obtained using a modification of a SMMR algorithm described by Wentz et al. (1986). Since, to our knowledge, neither the precise computational details of the algorithm (e.g., model coefficient values for the SSM/I) nor the results of any detailed validation studies have yet been published, we are not in a position to discuss its performance relative to those algorithms described earlier or relative to our own algorithms.

3. Algorithms

a. Surface Wind Speed

For near surface wind speed U , the Global D-Matrix algorithm of Goodberlet et al. (1989; hereafter referred to as GSW) remains attractive for its simplicity and apparent reliability under most conditions, provided care is taken to exclude pixels contaminated by precipitation. The algorithm is given by

$$U(\text{m s}^{-1}) = 1.0969T_{19V} - .4555T_{22V} - 1.760T_{37V} + .7860T_{37H} + 147.90 \quad (1)$$

where the subscripts for each SSM/I brightness temperature T indicate frequency and polarization.

In a minor departure from GSW, we flag U as indeterminate in pixels for which $T_{19V} > 215 \text{ K}$ or $T_{37V} > 221 \text{ K}$. In terms of atmospheric opacity due to rain, these thresholds are essentially identical, under typical conditions, to the Rain Flag "0" criteria specified by GSW; we have substituted vertically-polarized 19 and 37 GHz brightness temperature thresholds simply in order to reduce the sensitivity of the rain test to wind induced surface roughness. In a further modification to the original GSW algorithm, our experience to date has indicated that it is highly advisable to discard not only those pixels flagged as described above, but adjacent pixels as well. We surmise that this is because

of the tendency for the low resolution 19 GHz channels to be significantly affected by precipitation in neighboring pixels, even when the higher resolution channels are not.

After rejecting all pixels suspected of contamination, we use a (3×3) -pixel moving average to smooth the retrieved U field. This operation significantly reduces pixel-scale noise, while having little effect on the desired synoptic scale wind field. Following smoothing, remaining areas of missing values are "patched" by interpolating values inward from the valid pixels bordering the missing regions. This step is taken primarily to provide reasonable estimates of U for use in the retrieval of the other variables.

b. Integrated Water Vapor

In deriving our own statistical algorithm for integrated water vapor, we used the same radiosonde/SSM/I data set employed by Alishouse et al. (1990), but with slightly modified criteria for selecting pixels to be included in the regression. However, we chose to construct the algorithm from only the three lowest frequency channels of the SSM/I (T_{19V} , T_{19H} , and T_{22V}) so as to minimize the impact of cloud and precipitation on the retrievals. (For other combinations of channels, atmospheric liquid water not only has higher opacity, but large differences in surface resolution imply that the fractional coverage by cloud or rain could be quite different for each channel used in the algorithm.) Furthermore, we used $\log(280 - T_B)$ as the independent variable for each channel, this function being more nearly linear with respect to atmospheric opacity than the untransformed brightness temperatures.

The resulting regression equation, based on 515 nearly coincident soundings, exhibits overall error statistics (outside of precipitation) virtually identical to the Alishouse et al. (1990) algorithm applied to the same dataset, but with significantly reduced biases at very low and very high water vapor values (Fig. 2). RMS difference between our algorithm and the radiosonde values is the greater of 2.0 kg m^{-2} or $0.13V$. Of course, an unknown fraction of this difference arises from errors in the soundings themselves, as well as from spatial and temporal mismatches.

Our water vapor algorithm is given by the following:

$$\begin{aligned} V(\text{kg m}^{-2}) = & 11.98 \log_e(280 - T_{19V}) \\ & + 42.06 \log_e(280 - T_{19H}) \\ & - 54.36 \log_e(280 - T_{22V}) - 20.5. \end{aligned} \quad (2)$$

In practice, contamination of V estimates by rain becomes noticeable when $(T_{19V} - T_{19H}) < 24 \text{ K}$. Using this test as a criterion to flag contaminated V , we smooth and "patch" the retrieved V field as described above for U .

c. 85 GHz precipitation estimates (scattering based)

Spencer et al. (1989) describe a general technique for detecting large ice particles in rain clouds using a "polarization corrected" 85 GHz brightness temperature PCT which responds to volume scattering by these particles. This variable appears to be strongly correlated with surface precipitation intensity, particularly in convection and heavier frontal rain bands. However, as formulated by Spencer et al., who used constant linear coefficients for the two 85 GHz channels, significant large-scale variations in background PCT arise due to changes in atmospheric and surface temperature and surface wind roughening.

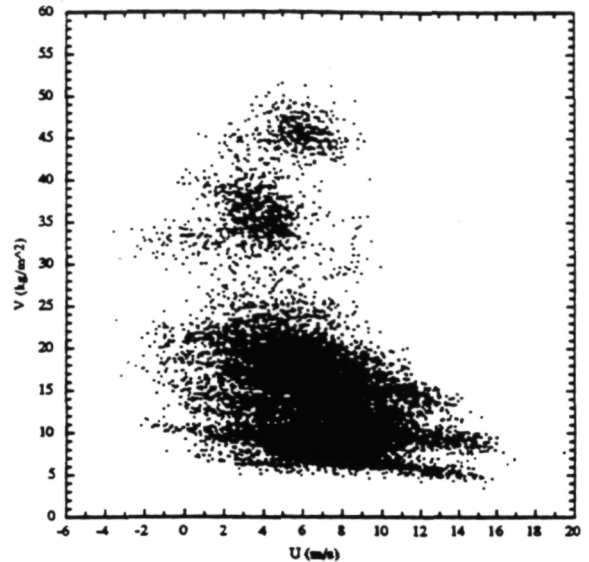


Fig. 3 Distribution of U and V for 12,287 pixels used to estimate dependence of SSM/I brightness temperatures and polarizations on U and V under cloud-free conditions and to test cloud water algorithms. U and V were estimated using the algorithms described in this paper. Pixels were obtained primarily from the Eastern Pacific and Gulf of Mexico; these were identified as essentially cloud free on the basis of GOES visible imagery. Note that the GSW algorithm for U occasionally returns negative wind speeds.

Here we propose an alternative formulation of the volume scattering signal which, while otherwise conceptually identical to that of Spencer et al. (1989), may be interpreted as allowing the slope of the assumed "pure emission line" (as seen in a bipolarization plot of 85 GHz brightness temperature) to vary in such a way that the latter always passes through (or near) the true "open ocean" brightness temperature pair and through a fixed, unpolarized reference brightness temperature corresponding to a opaque, non-scattering cloud at the freezing level.

We thus calculate the polarization corrected brightness temperature depression S as

$$S(K) = P_{85} T_{85V,O} + (1 - P_{85}) T_C - T_{85V}, \quad (3)$$

where the normalized 85 GHz polarization P_{85} is the ratio of the observed polarization difference Δ_{85} to the polarization difference $\Delta_{85,O}$ expected in the absence of clouds under otherwise similar conditions; similarly, $T_{85V,O}$ is the value of T_{85V} expected under cloud-free conditions. $T_C (= S + PCT)$ is the limiting brightness temperature of an increasingly opaque, non-scattering cloud layer, which we take to be a constant 273 K.

Expressions for $\Delta_{85,O}$ and $T_{85V,O}$ were found by regressing observed Δ_{85} and T_{85V} against retrieved U and V for 1.2×10^4 SSM/I pixels obtained from known cloud-free regions of the ocean (Fig. 3). This led to the following expressions:

$$P_{85} = (T_{85V} - T_{85H}) \exp(.0241U + .0271V - 4.44), \quad (4)$$

$$T_{85V,O} = 280.0 - \exp(4.20 - .00567U - .0406V). \quad (5)$$

In practice, we have found that S consistently remains within 10 K of zero except in precipitation. When $S > 10$ K, a very limited and preliminary set of comparisons with digital radar observations of light to moderate midlatitude frontal rain suggest that the surface rain rate R is given to within $\pm 1 \text{ mm hr}^{-1}$ by

$$R_s(\text{mm hr}^{-1}) \approx 0.25(S - 10), \quad (6)$$

The coefficient appearing in (6) will almost certainly be adjusted as further comparisons are conducted using more extensive radar comparison data. (Note also that while $S > 10$ appears to provide reliable evidence for the presence of precipitation, $S < 10$ is *not* sufficient to rule out light stratiform precipitation.)

d. Cloud Liquid Water

In the absence of anisotropic extinction by precipitation particles, it can be shown (e.g., Petty, 1990) that the normalized polarization P at any SSM/I frequency is well approximated by the effective microwave transmittance τ_C associated with liquid water alone, raised to a power α (≈ 2) which depends only weakly on other variables. That is,

$$P \approx \tau_C^\alpha \approx \exp \left[-\frac{\alpha \kappa L}{\cos \theta} \right], \quad (7)$$

where κ is the effective mass extinction coefficient of the liquid water in the cloud, L is the vertically integrated liquid water mass, and θ ($\approx 53^\circ$) is the SSM/I viewing angle.

Of all the SSM/I channels, those at 85 GHz have the best spatial resolution and sampling density and correspond to the highest κ for liquid water. We used the analytic brightness temperature model of Petty (1990) to determine $\alpha \kappa$ at this frequency, assuming conditions typical for regions of marine stratus near the California coast. We then inverted (7) to obtain

$$L_{85}(\text{kg m}^{-2}) = -0.339 \log_e(P_{85}), \quad (8)$$

where P_{85} is given by (4).

When significant radiative scattering by preferentially oriented, precipitation-size ice particles occurs, the 85 GHz polarization difference is no longer a useful measure of cloud opacity. Consequently, we flag L_{85} as indeterminate whenever S (see previous subsection) exceeds 10 K.

The scaling coefficient shown in (8) is strictly correct only for a cloud temperature near 285 K; for colder or warmer clouds there will be systematic relative errors ($\sim 10\%$ or more) in retrievals of L using this formula, owing to the strong dependence of κ on temperature. For marine stratus, of course, the temperature of the cloud is normally confined to a fairly narrow and well-determined range because of the strong coupling between cloud temperature and ocean surface temperature.

For most other weather regimes, however, cloud temperature may be extremely difficult to specify *a priori*, and a variety of other problems (e.g., precipitation, strong horizontal inhomogeneity, strong water vapor emission in the tropics) may also adversely affect the absolute accuracy of

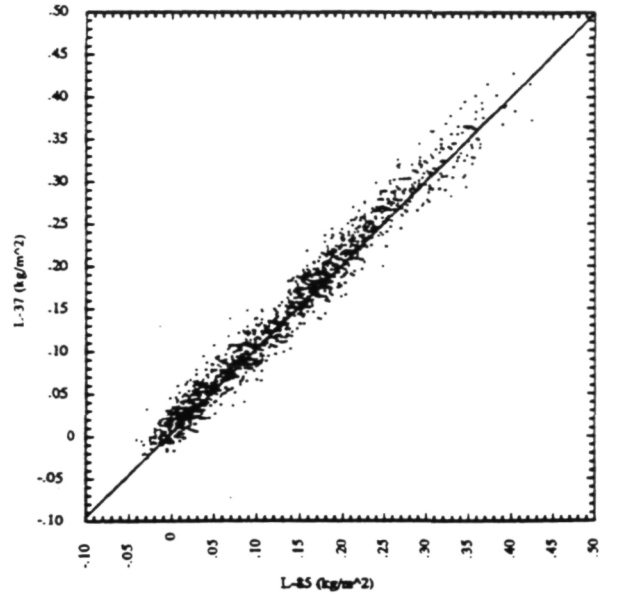


Fig. 4 SSM/I estimates of integrated cloud liquid water L obtained from the 85 GHz channels (Eq. 8) vs. those obtained from the 37 GHz channels (Eq. 9), for a region of marine stratus off the coast of Southern California. Data were obtained from three successive overpasses of the area, at 1336 UTC 16 July, 0130 UTC 17 July, and 1324 UTC 17 July, 1987. Solid line represents least-squares fit, with slope 0.99 and intercept 0.004. 85 GHz L was averaged over a 3 by 3 pixel box in order to approximate lower spatial resolution of 37 GHz estimates.

cloud water estimates. For example, if precipitation size droplets contribute a significant fraction of the total water mass of the cloud, (8) will give an *overestimate* of L owing to greatly enhanced (i.e., factor 2–4 larger) mass extinction coefficients of droplets with radii between 0.15 and 1.5 mm. Also, if there are inhomogeneities of order 0.3 kg m^{-2} or greater within the 85 GHz footprint (as might be the case with cold air mass cellular convection), the non-linear relationship between P_{85} and L may lead to significant *underestimates* of L from (8).

Consequently, there is probably little to be gained by using a more complex expression for L_{85} which would explicitly include cloud temperature.

A relation analogous to (8) holds between integrated cloud water L and P_{37} . Again using the theoretical brightness temperature model to determine $\alpha \kappa$, we have the following alternate algorithm for L :

$$L_{37}(\text{kg m}^{-2}) = -1.42 \log_e(P_{37}). \quad (9)$$

where the same approach as before was used to derive an expression for P_{37} :

$$P_{37} = (T_{37V} - T_{37H}) \exp(.0151U + .00607V - 4.40), \quad (10)$$

Advantages of L_{37} over L_{85} include reduced sensitivity to errors in V and to precipitation size particles, as well as greater linearity in the response of 37 GHz channels to liquid water (hence reduced errors due to inhomogeneity within the satellite footprint). Disadvantages include significantly poorer spatial resolution (30 km vs. 15 km) and, in middle and high latitudes (i.e., for low V), poorer signal-to-noise ratio.

We applied the above algorithms for L_{37} and L_{85} to our cloud-free data set and verified that the mean retrieved value of L was essentially zero; standard deviations were 0.025 kg m^{-2} and 0.021 kg m^{-2} , respectively. We also applied both algorithms to a region of marine stratus sampled on three successive passes near the coast of Southern California, and we compared the retrieved values of L (Fig. 4). Remarkably, the slope of the best fit line for these data is 0.99; this confirms that at least the ratio of the theoretically derived coefficients appearing in (8) and (9) is correct for marine stratus conditions.

e. 37 GHz precipitation estimates (attenuation based)

Petty and Katsaros (1990a,b,c) discuss the use of the normalized 37 GHz polarization difference P_{37} (see above) as an index of precipitation intensity and/or fractional coverage by precipitation. The results of those studies, which employed data from the older Scanning Multichannel Microwave Radiometer (SMMR), suggest that $P_{37} < 0.8$ ordinarily occurs only in the presence of precipitation (mean R increasing as P_{37} decreases to near zero), while $P_{37} > 0.9$ generally implies precipitation-free pixels. Values between 0.8 and 0.9 imply at most light precipitation, if any.

Pending the results of comparisons between surface radar data and P_{37} obtained from the newer SSM/I, we tentatively assume that the SMMR thresholds described above are also valid for the SSM/I, both in the tropics and midlatitudes. See Petty and Katsaros (1990b,c) for a discussion of the interpretation of P_{37} in terms of actual footprint averaged rain rate.

4. Summary

A unified set of algorithms has been presented for obtaining fields of surface wind speed, integrated water vapor, cloud liquid water, and precipitation from SSM/I brightness temperature data over the ocean. A graphic display and analysis package which implements these algorithms has already seen extensive use at the University of Washington for the study of the structure of oceanic weather systems.

Unique features of the algorithms described here include: (1) a semi-physical method for retrieving integrated cloud liquid water at the resolution of the 85 GHz channels, with no appreciable absolute bias and with rms absolute errors (as measure in cloud-free conditions) of only $\sim 0.02 \text{ kg m}^{-2}$; (2) essentially unbiased estimates of integrated water vapor over the entire natural range of this parameter, using a three-channel statistical algorithm which was constructed to have minimum sensitivity to precipitation; and (3) two complementary indices of precipitation activity (based on 37 GHz attenuation and 85 GHz scattering, respectively), each of which are constructed so as to be relatively insensitive to variations in other environmental parameters.

It is recognized that those SSM/I algorithms which depend on the 85 GHz vertically polarized channel cannot be used for data obtained from the F-8 SSM/I after mid-1988; however, nearly a complete year's worth of SSM/I data is available prior to that time, and a series of new SSM/Is, presumably with all seven channels functioning, are scheduled to begin replacing the F-8 sensor in the near future.

Acknowledgments

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Legacy of the Seasat Mission for Studies of the Atmosphere and Air-Sea-Ice Interactions

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Abstract

The polar-orbiting satellite, Seasat, had been designed as an oceanographic satellite with little advance thought being given to atmospheric uses. However, the microwave instruments provided a rich source of data for studying atmospheric conditions. The simultaneous sampling by several instruments generated special benefits, a situation which to date has not been repeated. In this review we emphasize studies of midlatitude and tropical cyclones and regional weather and climate analyses. We also touch upon studies of long swell, sea ice, and continental ice sheets with Seasat data.

Many of these results of the Seasat mission were serendipitous. In preparation for the major NASA initiative for the next decade, the Earth Observing Satellite (EOS) program, we thought it timely to bring some of the Seasat experiences to the fore, since valuable lessons can be learned from the successes and the failures (or omissions) of the Seasat program. We have learned of: 1) the synergistic value of integrated, overlapping sampling by several instruments, 2) the invaluable contribution of carefully planned surface measurements, and 3) the importance of retaining flexibility in the system (enough data retention) to allow unexpected and innovative analysis techniques.

1. Introduction

The Seasat satellite, launched in 1978, was originally advocated as a satellite for oceanographic research. The benefits that the Seasat mission has brought to the science of oceanography were reviewed by Stewart (1988). In addition, the suite of sensors on Seasat brought significant advances in our understanding of the atmosphere and provided new tools for diagnosing weather systems over the ocean. These results were largely serendipitous, but they have had far-reaching consequences in atmospheric sciences. The Seasat sensor design was derived from several prior instruments in space that provided microwave sensing of atmospheric water vapor, cloud liquid water, and sea ice. This atmospheric and surface information over the global oceans may be particularly important, since quantitative measurements over wide expanses of the ocean are still very limited, especially in the Southern Hemisphere. The many different sensors operating concurrently on the Seasat satellite also provided a unique view of certain geophysical phenomena, such

as cyclonic storms. The ice covered ocean, continental ice sheets, and long ocean surface waves could also be monitored from space.

Visible and infrared imagery from space has often provided the main clues about weather in regions distant from meteorological stations. This information has greatly improved weather analysis and forecasting. However, it is still severely limited information. The intensity of reflected sunlight can sometimes give information about the thickness of clouds, but the lower atmosphere or the sea surface is very often completely obscured by clouds. Infrared signals emerging from the uppermost regions of most clouds give no information about what exists below. Data from more complex satellite instruments such as sounders are often contaminated by cloudiness. In contrast, certain frequencies of microwave radiometers receive an integrated measure of the atmospheric emission through the entire atmospheric column, even in cloudy conditions. The microwave radiometer on Seasat, the Scanning Multichannel Microwave Radiometer (SMMR), produced information on integrated atmospheric water vapor, cloud liquid water, and rain over the whole atmospheric column and could sense sea surface temperature and wind speed in cloudy conditions. The active radar signal from the scatterometer is sensed by microwave radiation reflected or scattered off the sea surface in most cloudy conditions. Hence, microwave sensors provide unprecedented information about a weather system which is complementary to the visual or infrared pictures.

There is a correlation between the surface roughness characteristics seen by microwaves and either the near surface wind or the surface stress (e.g., Jones et al. 1982; Brown et al. 1982; Thompson et al. 1981, 1983). For the oceanographic mission the surface wind or stress fields measured by the SASS (Seasat A Satellite Scatterometer) provide the upper boundary condition for studies of oceanic dynamics. At the same time, they provide the lower boundary condition for the atmospheric flow. This allows calculation of gradient flows, which can be interpreted in terms of the synoptic scale surface pressure gradient field. This is valuable meteorological information—not available by any other direct measurements for large regions of the ocean. Weather analysis in the South-

ern Hemisphere may be revolutionized by this capability.

There were high expectations for valuable data from the Seasat microwave instrument ensemble for the data-poor Arctic regions. Despite Seasat's brief existence (104 days), the key capabilities for observing sea ice have been demonstrated. The active signal from the SAR (Synthetic Aperture Radar) together with SMMR and SASS data yielded a synergistically valuable dataset for this purpose. Information on sea ice kinematics, sea ice types, the ice edge, topography of ice sheets, surface waves, and winds provided a broad new look at an otherwise almost inaccessible region.

Because major new satellite systems are currently being designed, it is important to develop an awareness of the potential value for meteorological and air-

was also derived from a prototype flown on Skylab and especially from an altimeter flown on the GEOS-3 satellite. The fourth microwave instrument was the synthetic aperture radar, SAR. Seasat also carried a visible and infrared radiometer VIRR, which only functioned well for the first month and was not as advanced as related instruments on the geostationary satellites and on the NOAA (National Oceanic and Atmospheric Administration) operational satellites.

a. The radiometer—SMMR

The Seasat SMMR measured five frequencies in both horizontal and vertical polarization (Gloersen and Barrath 1977; Njoku et al. 1980). It was intended to provide a correction to scatterometer wind results for cloud water, and to provide sea surface temperature data in the presence of clouds. The frequencies in GHz

and their nominal associated ground resolution in km are 6.6 (150), 10.4 (80), 18 (50), 22 (50), and 37 (25). Each frequency is particularly sensitive to certain atmospheric and sea surface parameters. The quantities of interest here are listed in Table 1. The algorithms for deducing these variables usually involve several

Because major new satellite systems are currently being designed, it is important to develop an awareness of the potential value for meteorological and air-sea-ice interaction research and operations that are offered by passive and active microwave sensors.

sea-ice interaction research and operations that are offered by passive and active microwave sensors. There is great opportunity for enhancement of weather, climate, air-sea interaction, and sea ice information when many different sensors are used in conjunction. Careful planning will make these future satellite-borne systems much more valuable, and the Seasat legacy, illustrated in the following examples, can aid us in this venture. We refer the reader to Stewart's (1988) article for the history of Seasat and general background information. The Seasat legacy has already influenced analysis of data from a few similar instruments operated in space during the 1980s such as Nimbus 7 SMMR, the Special Sensor Microwave Imager (SSM/I), and the Geosat altimeter. We will not review this work in detail here, but will only refer to it occasionally.

2. The Seasat microwave instruments: SMMR, SASS, ALT, and SAR

The Seasat satellite carried four microwave instruments. The SMMR was an offspring of instruments employed during a decade of satellite microwave radiometry. The SASS, an active microwave radar, was a new development. The only previous scatterometer flown in space was on Skylab, a short-term manned mission. The altimeter instrument, ALT,

frequencies, since corrections for attenuation or other modification of the signal by atmospheric effects must be included. For instance, cloud liquid water in the path must be accounted for when calculating integrated water vapor, and vice versa. The Seasat algorithms are described by Wentz (1983), Wilheit and Chang (1980), and Chester (1981). Sea surface temperature estimates require the lowest frequency (6.6 GHz) low resolution channel and have therefore not been of great interest, especially since infrared techniques that include corrections for atmospheric absorption and elimination of cloud-contaminated pixels have been much improved in the past decade. The Seasat algorithms were tested against intercomparison data collected in several field experiments (e.g., Lipes et al. 1979). The approximate accuracies of each of these variables are also listed in Table 1 with a column describing the intercomparison data and another with some relevant references. The most dependable and useful measure of the atmosphere provided by SMMR is undoubtedly the integrated water vapor in the column (IWV), which is of about the same accuracy as that obtained by integrating the humidity measured in radiosonde ascents (Katsaros et al. 1981; Alishouse 1983).

b. The scatterometer—SASS

SASS was designed to provide surface wind (or stress) vectors. This active radar signal operating at 14.6 GHz

TABLE 1: Geophysical parameters derived from SMMR and SASS and their accuracy.

		Accuracy	Comment	Reference
SMMR	Seasat integrated water vapor	$\pm 2-3 \text{ kg m}^2$	Has same bias and rms as integrated radiosonde values	Lipes et al. (1979) Alishouse (1983)
	Integrated cloud liquid water	unknown	Qualitative comparison with visible and infrared pictures is good	Katsaros et al. (1981) Alishouse (1983) McMurdie and Katsaros (1985) Katsaros and Lewis (1986)
	Rain rate	estimated factor of 2 to 3	Qualitative ship observations show the same trend as SMMR estimates	McMurdie and Katsaros (1985) McMurdie (1985)
	Frontal location	$\pm 50 \text{ km}$	Exact surface location of front ill-defined in comparison data	McMurdie and Katsaros (1985)
	Wind speed	$\pm 2 \text{ m s}^{-1}$	When rain and sunglint effects are removed	Cardone et al. (1983) Wentz et al. (1982, 1986) Lipes et al. (1979)
	Sea surface temperature	$\pm 2 \text{ K}$	The large footprint for SST makes the data useful only 300 km from land	Njoku et al. (1980) Bernstein and Morris (1983)
	Sea ice concentration	$\pm 5\%$	SAR measurements can improve accuracy	Swift et al. (1985)
	Sea ice multiyear concentration	$\pm 30\%$	SAR measurements can improve accuracy	Swift et al. (1985)
SASS	Wind speed	$\pm 2 \text{ m s}^{-1}$ $\pm 4 \text{ m s}^{-1}$	Synoptic averages Near fronts, storm centers	Jones et al. (1982) Brown et al. (1982)
	Direction	$\pm 20^\circ$ $\pm 40^\circ$	Synoptic averages Near fronts	
	Frontal location	$\pm 50 \text{ km}$	See comments for SMMR (above)	Brown (1983)

responds to the short gravity and capillary waves to produce a backscattered signal, which is called the "Normalized Radar Cross Section," σ_0 . The surface roughness signal could be correlated directly with surface stress, were such data available. Since stress is very difficult to measure, the signal has been correlated with surface wind measurements. The classic aerodynamic bulk parameterizations are then used to calculate estimates of surface stress, heat, and moisture flux from the wind data. However, this requires estimates of temperature and humidity near the surface, and knowledge of atmospheric stratification is important to these relations. The near-surface air temperature is the most difficult parameter to obtain

for these calculations. Despite great uncertainty in the physics of the relationship between radar backscatter signals, surface roughness features and near-surface winds, sufficient verification has emerged to lend confidence to the data product for a wide variety of uses (Table 1).

c. The altimeter—ALT

The Seasat 13.5-GHz altimeter (ALT) measured the nadir radar return in a narrow swath about 7 km along the subtrack. The return-trip travel time gives a measure of the satellite altitude. If the satellite's position is well known, the geoid can be mapped and temporal variations in the elevation of the sea surface can be

found. The main purpose of this instrument was to map surface signatures of ocean-bottom topography and determine ocean tides and currents (Stewart 1988). Interpretation of the shape and broadening of the return signal also allows determination of mean wave height ($H1/3$) and wind speed (Fedor and Brown 1982), since the small scale roughness of the sea responsible for the intensity of the radar return depends mainly on local wind speed. However, comparison of Seasat altimeter wind speeds to SMMR and SASS wind speeds (Wentz et al. 1982) showed significant differences. Later wind speed algorithms for altimeters have shown better agreement with in situ comparison data (e.g., Chelton and Wentz 1986; Mognard et al. 1984). Topography of continental ice sheets can also be mapped with altimeter data (e.g., Zwally et al. 1983; Remy et al. 1989).

One major advantage of satellite remote sensing for atmospheric science is that the same instrument can sample the global atmosphere over a short time interval as compared to conventional data.

d. The synthetic aperture radar—SAR

The synthetic aperture radar operating at 1.3 GHz was the first such instrument flown on a satellite. It produced high-resolution images (25 m) over a 100-km wide swath by observing the same spot on the earth from several positions of the satellite. This technique provides a synthetically enlarged antenna which results in the high resolution. Only small portions of the earth's surface were observed by this instrument because of its extremely high data rate. However, the data have provided a unique view of ocean waves, bottom topography, and sea ice dynamics.

SAR images show the common observation of wind streaks parallel to the wind from the helical roll circulation in the planetary boundary layer (PBL). These streaks are typically 1 to 3 km apart. This feature was used by Gerling (1986) to infer wind vectors from SAR data. He used SAR backscatter returns to estimate the wind speeds to about the same accuracy as for SASS, and then retrieved the wind direction from the orientation of the roll streaks. He estimated 10° precision in direction on a 6-km horizontal scale.

3. The global perspective

One major advantage of satellite remote sensing for atmospheric science is that the same instrument can

sample the global atmosphere over a short time interval as compared to conventional data. Much emphasis was spent on producing global maps of the surface wind field over the oceans on a 100-km scale and incorporating them into forecast schemes (Wurtele et al. 1982). Probably the best known Seasat accomplishment is our ability to produce global ocean-surface wind fields such as the example shown in Fig. 1. These wind fields have been used to initialize regional atmospheric numerical forecast models. In some cases, no improvement in the forecast was discerned until a vertical distribution of the influence from the unsmoothed, raw surface divergence field was prescribed (Duffy and Atlas 1986). This points to the importance of contemplating how satellite data are incorporated into conventional models, which have been adapted to handle the low-quality conventional data over the ocean. Such assimilation requires close collaboration between specialists in the related sub-fields of numerical modeling, remote sensing, and boundary layer studies.

Global monitoring of atmospheric water parameters from the Seasat SMMR was soon eclipsed by the much larger record from the Nimbus 7 SMMR (launched in October 1978 and operated until 1987, e.g., Gloersen et al. 1984). With this dataset, seasonal and interannual variability could be explored (e.g., Prabakara et al. 1982, 1983, 1985). However, while Nimbus 7 SMMR operated only on alternate days, the Seasat SMMR operated continuously, which was a distinct advantage for studying the evolution of storms crossing the oceans. Analysis of Seasat SMMR data also benefited from the overlap with scatterometer data (see sections 4 and 5 below).

Liu (1984) used Seasat SMMR data to estimate the turbulent water vapor flux from the ocean. He found that monthly averaged column-integrated water vapor could be used as a proxy for near-surface humidity, and the SMMR-produced sea surface temperature allowed estimates of interfacial humidity. Together with sea surface temperature data and wind speed from either SMMR or the scatterometer, the bulk aerodynamic formulas were then used to estimate evaporation rates (Liu and Niiler 1984). This method for estimating evaporation in the tropical Pacific was used with more recent microwave radiometer data from the Nimbus 7 satellite and provided insight into the roles played by water vapor flux and sea surface temperature changes in the El Niño phenomenon (Liu 1989).

Etcheto and Merlivat (1988) employed Seasat scatterometer wind data with a relationship between gas transfer velocity and wind (Liss and Merlivat 1986) to deduce the carbon dioxide exchange coefficient for the global ocean. They concluded that even though

**Gridded wind vector field from Seasat scatterometer winds
15-day mean, 6 September 1978–20 September 1978**

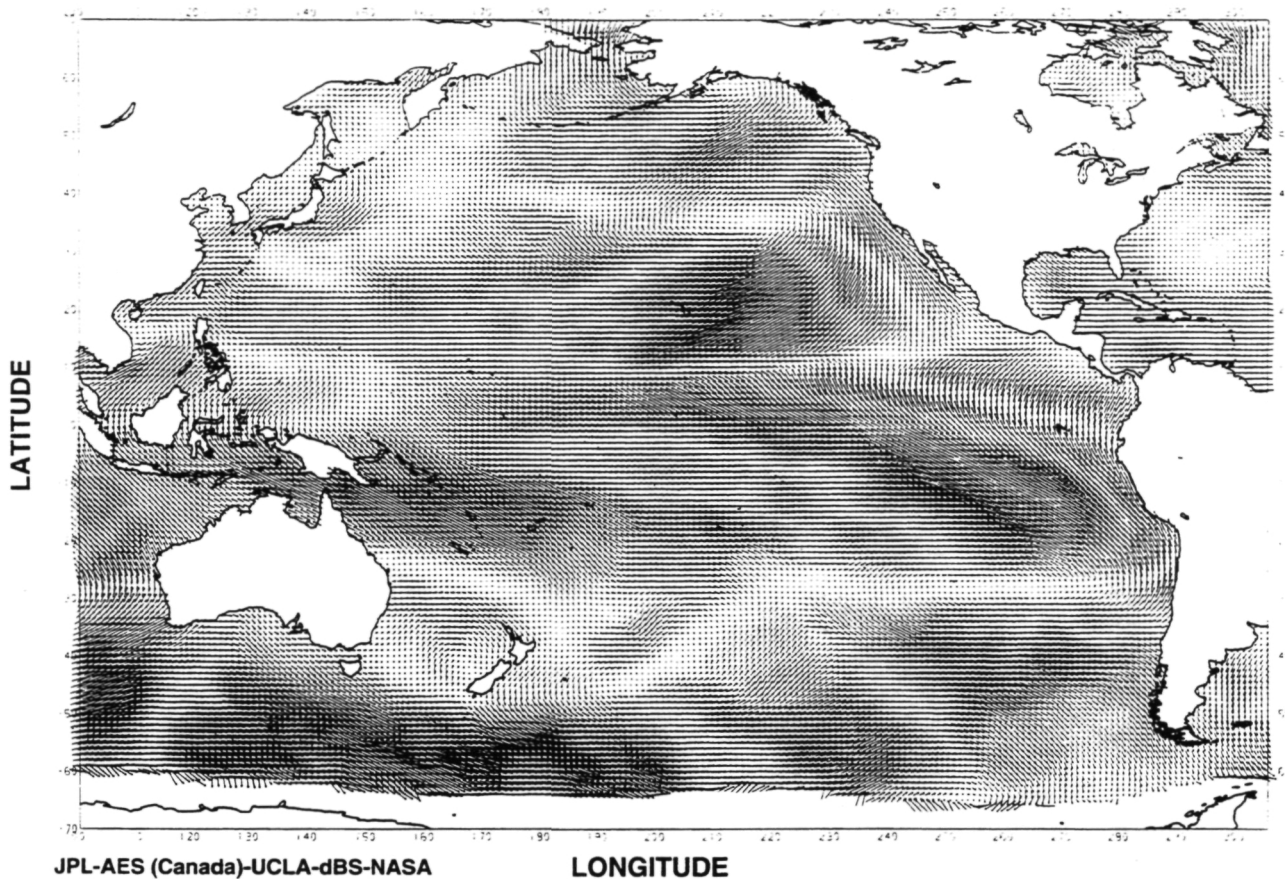


FIG. 1. Vector wind fields for the Pacific Ocean, from 15 days of SASS swath data, September 1978. (Fig. 1 was contributed by P.M. Woiceshyn and G.F. Cunningham, Jet Propulsion Lab; M.G. Wurtele, UCLA; S. Peteherych, Atmos. Env. Service, Canada; and D.H. Boggs, dB Systems 1990.)

short in duration, Seasat wind data provided a valuable means for a first attempt to estimate global carbon dioxide fluxes. Their result for the net carbon dioxide flux is within a factor of 2 to 3 of estimates for longer periods made by other means.

The SASS signal also produces the lower boundary condition for the atmospheric flow. With a planetary boundary layer (PBL) model (Brown and Liu 1982), the solution for the wind profile up to the gradient wind balance becomes available. This profile may depend significantly upon other factors, such as layer stratification and baroclinity (thermal wind), the determination of which requires additional data sources. When the auxiliary information is available from other satellite instruments, in situ measurements, or numerical models of the atmosphere, the global marine wind field from surface to gradient levels can be produced with high horizontal resolution.

The gradient wind fields can be integrated to produce pressure gradient fields, or if a benchmark pressure measurement is available, absolute pressure

fields can be generated (Brown and Levy 1986). There is excellent agreement with National Meteorological Center pressure fields in the regions of the Northern Hemisphere, where many ship and buoy data are available. This agreement, shown in Fig. 2, is strong additional evidence that the winds from a scatterometer are within the stated accuracy. Such surface pressure fields may provide a much needed correction for altimetry—the correction for the inverted barometer effect on sea level (R. H. Stewart, personal communication).

The SASS/PBL-analyzed pressure fields show some significant differences in the Southern Hemisphere. Figure 3a,b shows the correspondence between SASS/PBL-derived surface pressure fields and those produced by the Australian Bureau of Meteorology and the European Centre for Medium Range Weather Forecasts (ECMWF). The SASS/PBL derived pressure fields agree better with those produced by the numerical analysis in regions where the latter have a surface observation or two. However, as coverage

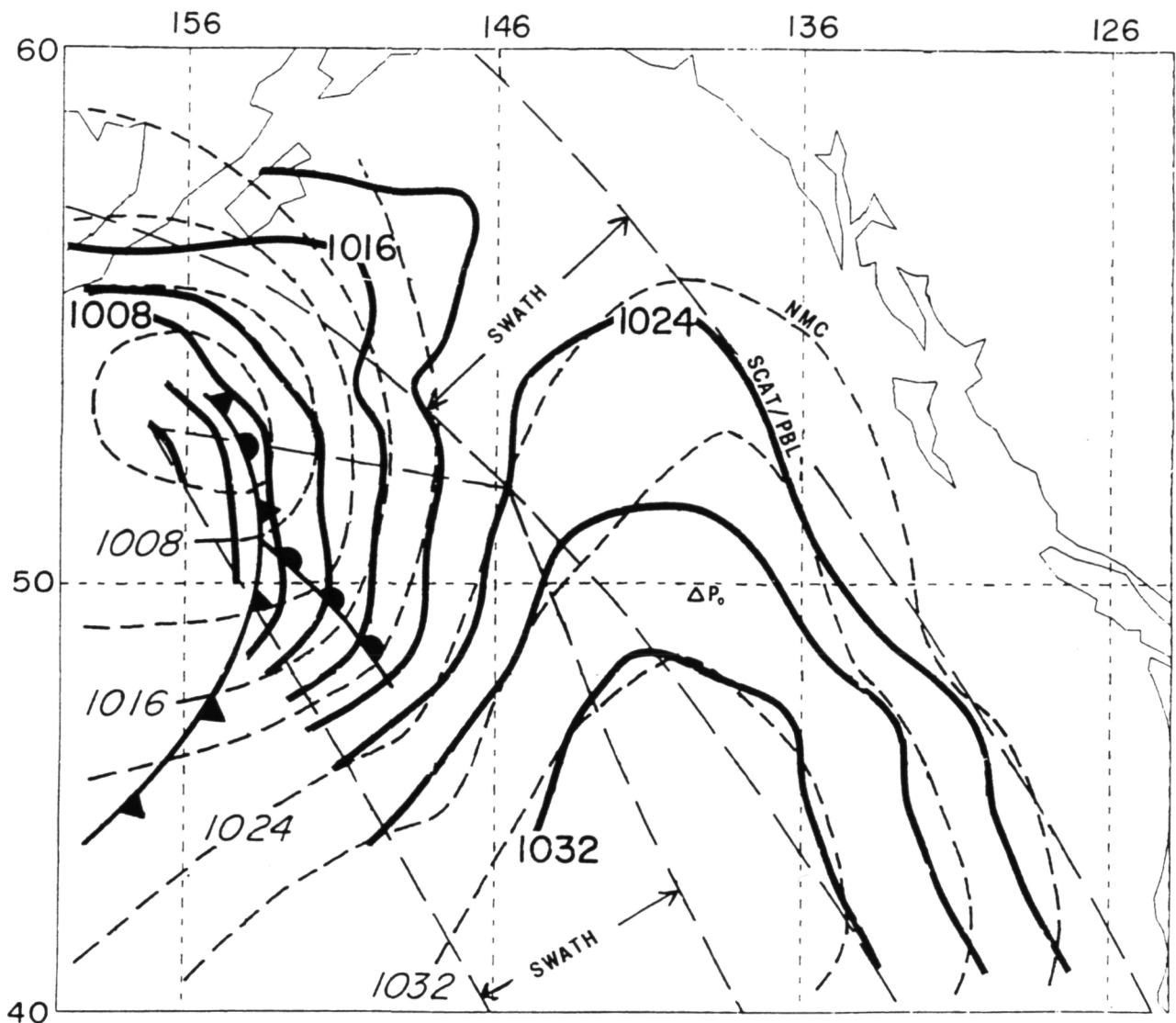


FIG. 2. A comparison of atmospheric surface pressure fields derived from SASS data using a PBL model and NMC data over North Pacific Ocean. SASS/PBL fields are solid lines and National Meteorological Center fields are dashed lines. Long dashes enclose satellite data swaths, and the triangle is a single surface pressure point used in the model (after Brown and Levy 1986).

continues toward Antarctica, there are generally large differences between the SASS-derived fields and the conventional fields. Also, there are often indications of mesoscale variations in the SASS/PBL data that are absent in the analyses of the weather services. The SASS/PBL procedure applied in the Southern Hemisphere invariably produces stronger pressure gradients; the lows are deeper (up to 30 mb), and the fronts are often relocated (Levy and Brown 1991). This occurs despite expectations that SASS winds would be underestimated at high wind speeds (Woiceshyn et al. 1986), and the use of a conservative, lower wind estimate in the SASS algorithm in areas of high wind speeds. The comparisons suggest strongly that weather service analyses in this region would benefit enormously from systematic scatterometer input. There

are enough similarities to anchor the fields in some areas, and enough differences to suggest that many of the weather and climate standards used to estimate global turbulent air-sea fluxes will have to be revised for this region.

A study of the relationship between surface wind speed, wind waves, and very long surface waves (swell) was carried out by Mognard et al. (1986). Because sea level varies by a substantial amount from crest to trough in the very large waves of the Southern Hemisphere, the swell can be well measured by the Seasat altimeter; very few direct observations of sea state are available from the remote and hostile regions of the Southern Ocean. Satellite altimeters such as ALT can therefore provide data for climatological as well as forecasting and warning purposes. Mognard

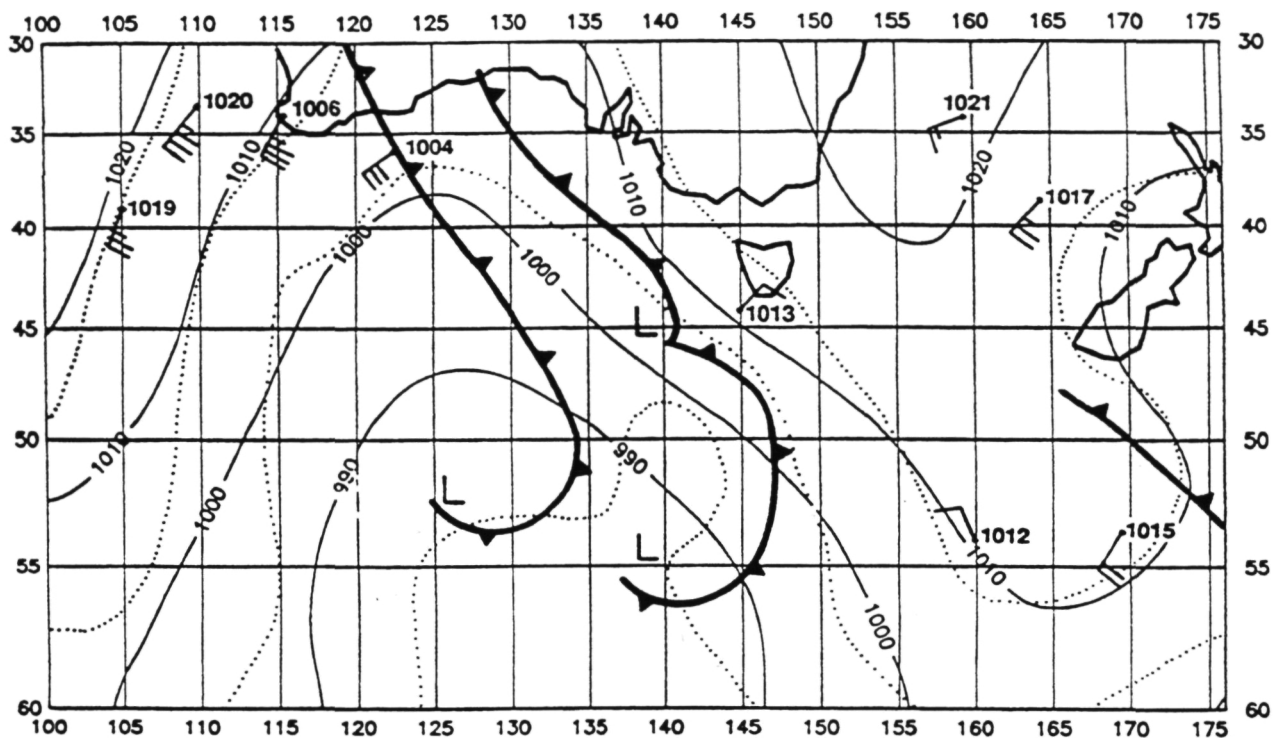


Fig. 3a. Comparison pressure fields from the Australian Bureau of Meteorology (solid) and the European Centre for Medium Range Weather Forecasts (dotted) for the Indian Ocean, 10 September 1978. Surface wind reports are shown. Frontal systems are derived from satellite cloud observation by ABM.

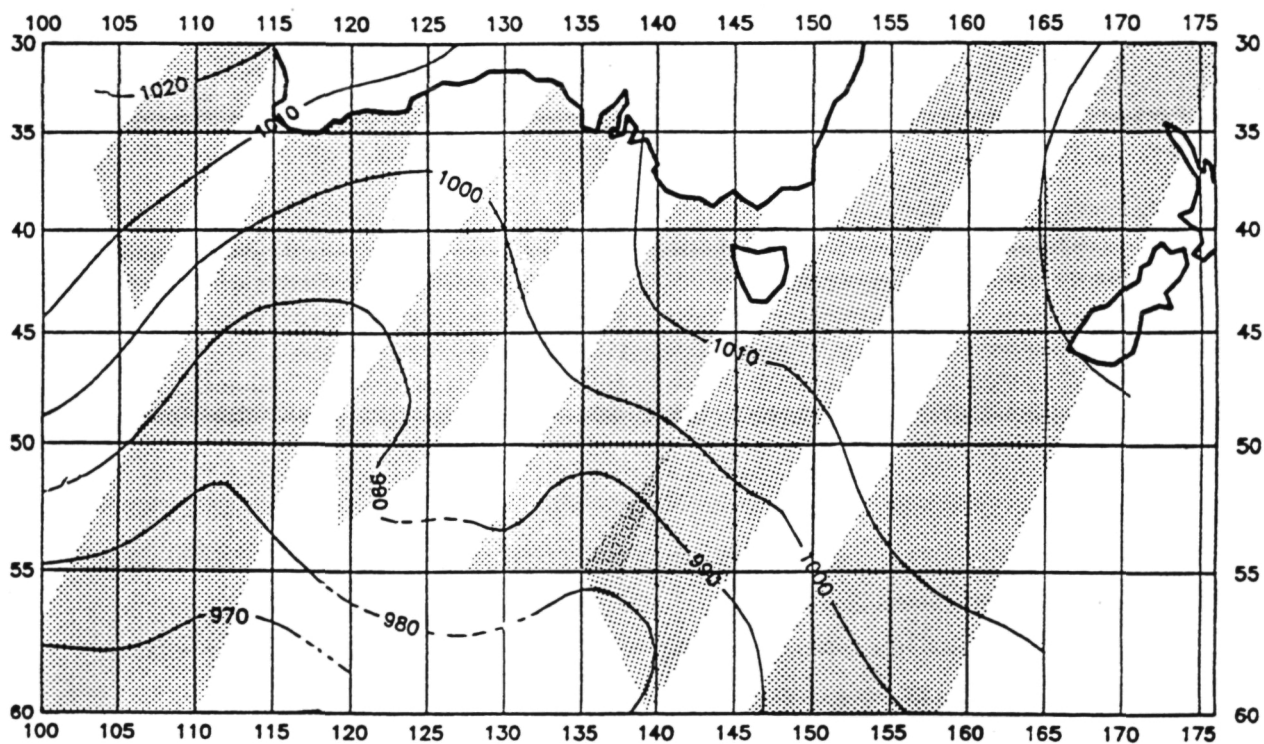


FIG. 3b. Southern Hemisphere pressure field from a SASS/PBL model. Shaded areas are SASS swaths from three orbits (± 90 min). Pressures are similar to FIG. 3a near Australia and differ significantly farther south.

(1984) was able to identify the long swells (wavelength of the order of 500 m) for many days as they travelled from the southern Pacific Ocean, across the equator, and up the Central American coast (Fig. 4).

4. Studies of midlatitude cyclones

SMMR

The maps of integrated water vapor which the Seasat SMMR provide give an instantaneous view of the areal distribution of water vapor in large regions of cyclonic weather systems. The Norwegian cyclone model provided the concepts of cold, warm, and occluded fronts,

Information obtainable from satellite-borne microwave remote sensors can be used to study detailed mesoscale features in midlatitude cyclones at scales of tens to hundreds of kilometers, and is unique and complementary to the synoptic scales seen in visible or infrared imagery.

and ideas for how the different air masses on either side of the fronts are arranged. Our understanding of the processes by which these contrasts develop is currently going through revisions, but the main features remain and can now be readily observed. Maps of integrated water vapor from Seasat's SMMR and later microwave radiometers provide an instantaneous picture of the contrasting air masses. A sketch of what the fields of integrated water vapor would look like with ideal sampling by a microwave radiometer during three stages of a cyclone's development is seen in Fig. 5. A similar picture had emerged from sequences of radiosonde ascents collected as overland frontal systems passed a station. These studies led to the concept that the region ahead of a cold front is a conveyor belt for warm, moist subtropical air. Certainly the picture in Fig. 5 is consistent with this general idea. This sketch was based on the cyclones studied with data from Seasat SMMR by McMurdie and Katsaros (1985) as well as our later work. A case from the North Pacific showing a mesoscale wave at 48 N and 146 W is seen in Figure 6a. Such features are not as readily identified in other forms of data, although such a region can sometimes be identified by extra bright or cold clouds in visible or infrared satellite images. Hobbs and Persson (1982) discuss many mesoscale

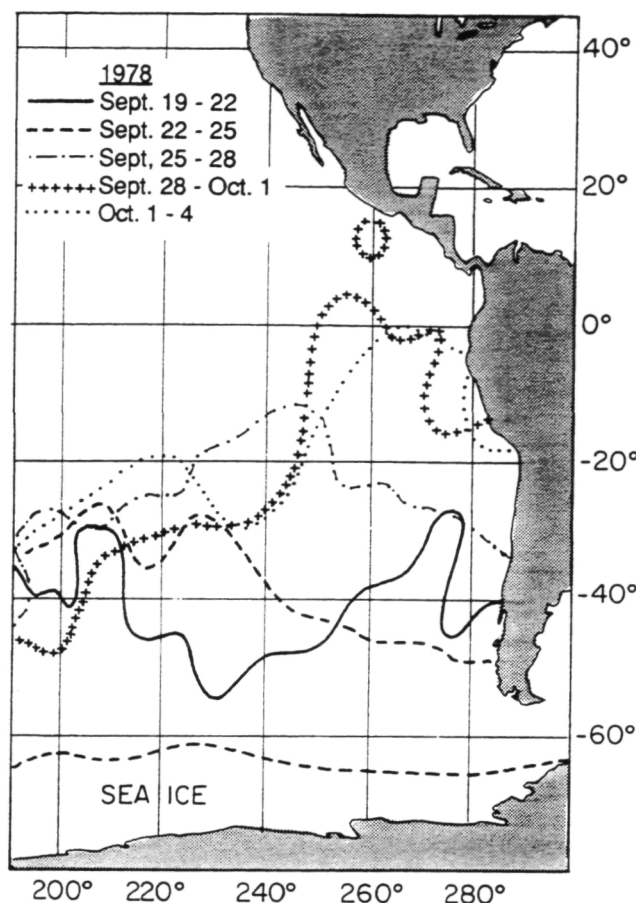


Fig. 4. Evolution of the isopleths of the 3-m significant swell height between 19 September and 4 October, deduced from the SEASAT altimeter data. Notice the rapid north-northeast progression from 22 to 30 September and the regression toward the southeast between 1 and 4 October (after Mognard 1984).

features of cold fronts which can be readily identified in microwave satellite images (see also Katsaros and Lewis 1986, who based their observations on data from the sister instrument to the Seasat SMMR on the Nimbus 7 satellite).

Information obtainable from satellite-borne microwave remote sensors can be used to study detailed mesoscale features in midlatitude cyclones at scales of tens to hundreds of kilometers, and is unique and complementary to the synoptic scales seen in visible or infrared imagery. Recent studies by McMurdie (1989) employing data from the FGGE (First GARP Global Experiment, where GARP stands for Global Atmospheric Research Program) produced at the ECMWF have shown that the SMMR-produced patterns in IWV can be attributed primarily to horizontal advection in midlatitude cyclones. The absolute amounts of IWV on the cold (dry) and warm (moist) sides of these fronts vary significantly from season to season and between different regions (McMurdie and Katsaros 1991).

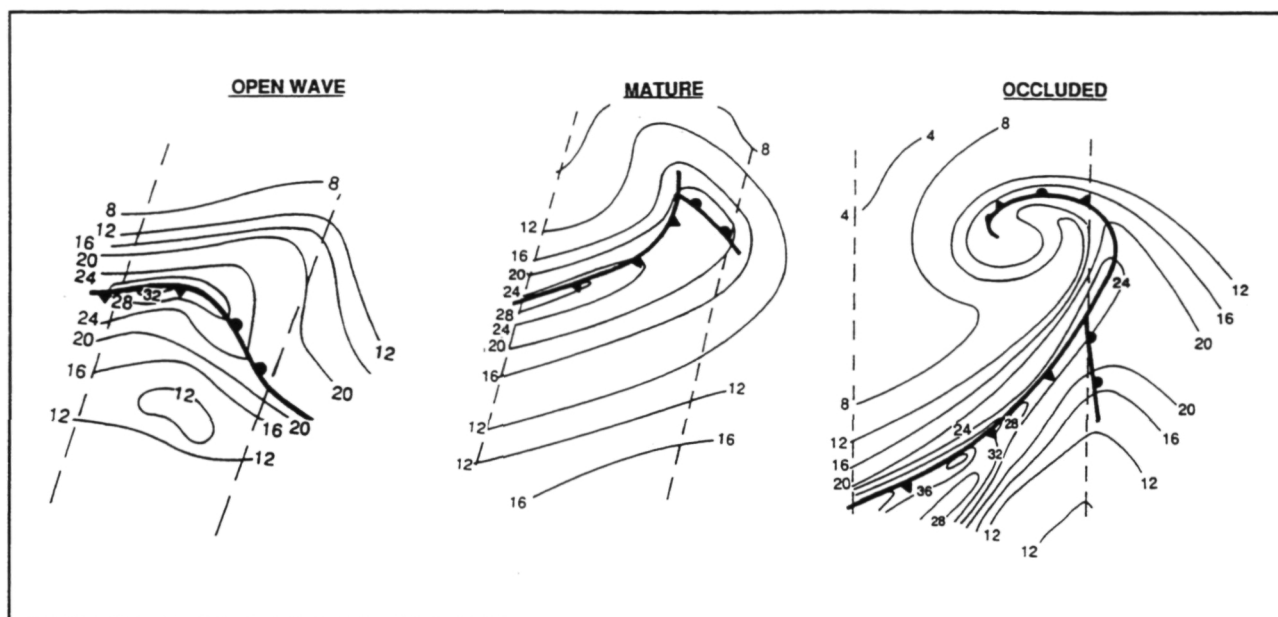


FIG. 5. Integrated water vapor in midlatitude cyclones—schematic distribution based on microwave radiometer data. Units are kg m^{-2} . The dashed lines are about 1300 km apart and indicate swath width of the modern microwave radiometer, SSM/I (Special Sensor Microwave/Imager). The narrower swath, SMMR on Seasat and Nimbus 7 is seen in Fig. 6. (The limited swath width somewhat hampered studies of cyclones with Seasat data.)

The pattern of IWV in midlatitude cyclones was found to be very regular as more cases were studied with Seasat SMMR data. The fronts identified by conventional techniques were consistently found to be located at the leading edge of the strong gradient in IWV, as seen in Figs. 5 and 6. In fact, as more data accumulated from the Nimbus 7 SMMR, it became clear that the gradient in IWV might be a good indicator of frontal location. In testing this hypothesis it was found that the threshold of the gradient, which would locate the fronts, was independent of both season and the region of the global ocean where the storms were found (Katsaros et al. 1989).

During the first Seasat data workshop (January 1979), one of us (K. Katsaros) had the thrilling experience of finding characteristic raincell and rainband structure in a North Pacific cyclone revealed in the patterns of rain rate [obtained by the still completely untested algorithms of Wilheit and Chang (1980) or Wentz (1983)]. Such structures were concurrently being discovered by radar and aircraft (Hobbs 1978) but had not been seen on such large scales before. The Seasat SMMR algorithm was later modified (Chester 1981), and several other algorithms for the SMMR frequencies (37 GHz in particular) have emerged (Spencer 1986; Petty and Katsaros 1990a,b,c). It was clear already in the study by McMurdie and Katsaros (1985) from the Seasat SMMR data that the low rain rate intensities determined in rather widespread cloud systems were in agreement

with the qualitative ship observations reported in the synoptic surface code.

SASS

With its ability to penetrate nonprecipitating clouds, the SASS provides data for studying mesoscale surface wind distributions in midlatitude storms (Brown 1983, 1986; Pierson 1983). The SASS signal characteristics can be used for locating fronts because the character of the backscatter signal varies dramatically across the front. This is due to a change in the amplitude or density of the small-scale surface waves, which is related to a change in surface stress or near-surface wind speed (e.g., Donelan and Pierson 1987). Geernaert et al. (1986) and Byrne (1982) discuss variations in the drag coefficient across fronts based on surface or aircraft measurements. Peteheych et al. (1988) found that adding scatterometer surface winds to the synoptic data improved short-range weather forecasts for the west coast of Canada and the United States in six case studies due to significant improvements in the surface analyses.

The SASS winds can be used to calculate the divergence and vorticity fields, which are basic quantities in synoptic weather and climate analyses. The nearest points to a front where these variables can be calculated with confidence from SASS data are approximately 100 km to each side. However, this is sufficient to show changes in these fields across the

**Comparison of SMMR water vapor patterns and precipitation to SASS divergence field
in a North Pacific cyclone at 1800 UTC 11 September 1978.**

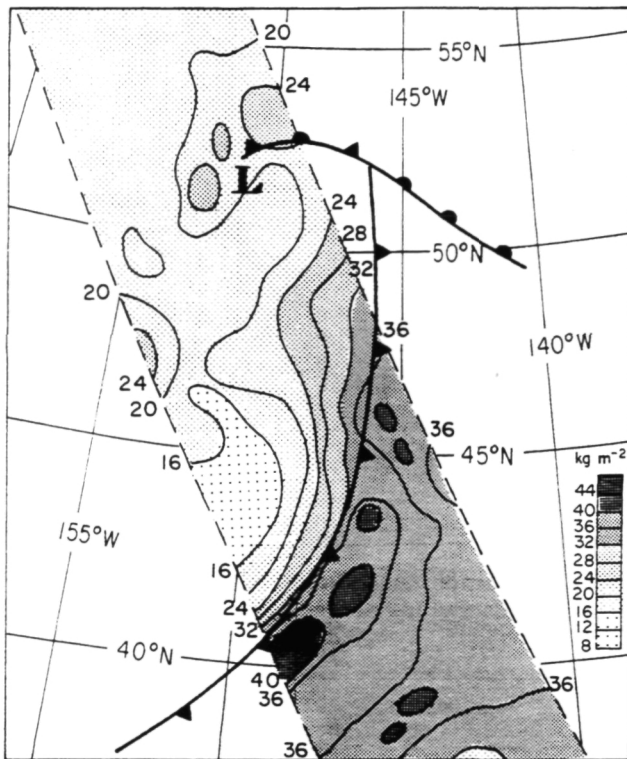


FIG. 6a. Integrated water vapor. Superimposed frontal analysis by McMurdie et al. (1987). The dashed lines indicate SMMR swath width.

front, which yields information on frontogenesis (Levy 1989). Scatterometer winds were also found to be valuable in hindcasts of surface waves (Janssen et al. 1989).

SASS and SMMR

The Seasat satellite is unique in satellite history to date in having overlapping microwave radiometer and scatterometer swaths (a wide-swath microwave radiometer and a scatterometer will not be on the same satellite, according to current plans, until 1998 or later). The divergence fields calculated with SASS winds have been compared to fields of SMMR integrated water vapor, yielding confidence in the adequacy of both parameters and providing more detailed information about cyclonic weather systems. McMurdie et al. (1987) report on several case studies where mesoscale features, such as the wave on the front seen in Fig. 6a by SMMR, is also identifiable in the SASS-derived convergence (Fig. 6b). Near frontal systems, surface convergence, and large IWV values have been found to be strongly correlated in the Seasat data. Occasionally, there are inconsistencies which require more data to resolve in terms of various

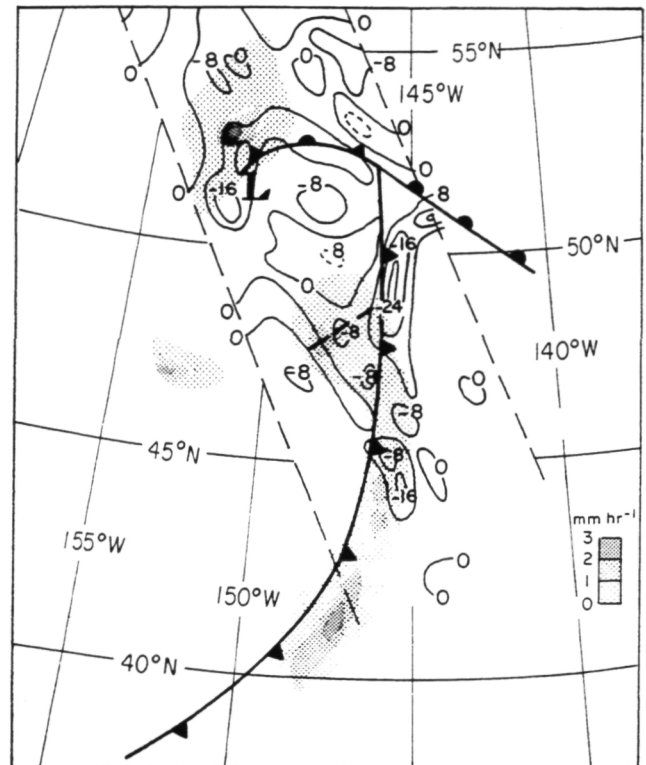


FIG. 6b. Divergence pattern of surface wind field in units of 10^{-5} s^{-1} derived from SASS with a modified frontal analysis also showing a line of convergence (heavy dashed line). The dashed lines indicate SASS swath width. The SMMR-derived precipitation rate in mm h^{-1} is also indicated with three grades of shading (after McMurdie et al. 1987 - see original article for details). Note that rain coincides with convergences where the two swaths overlap.

possible geophysical interpretations or simply in terms of system errors. For instance, when surface convergence is present but without a large amount of IWV, it might mean that the air is not picking up moisture because the surface air is saturated, the boundary layer flux is reduced by stable stratification, or the surface convergence has not been operating over a sufficiently long time interval. Data on the moisture flux and sequential data from the same region will resolve such questions. Where there is a large amount of IWV in the column but weak surface convergence, the IWV may be interpreted as being produced by upper-level convergence, horizontal advection, or simply evaporation from warm surface water over a period of time. The Seasat SASS and SMMR have raised many such questions in their brief tenure. In the future, when a scatterometer and a radiometer can sample cyclones together for an extended period of time, we can expect to learn much about the structure, dynamics, and evolution of cyclones. For reasonable temporal resolution of the evolution of cyclones, we should have two

polar-orbiting platforms, which would allow four views of most midlatitude regions per day.

5. Studies of hurricanes

Despite the short duration of Seasat, ten tropical storms were sampled, seven of which reached hurricane force. The symbiotic benefits of simultaneous analyses of data from SMMR, ALT, SAR, and SASS were shown in an analysis of Hurricane Iva by Gonzalez et al. (1982). Mean wave heights were derived from ALT, dominant wave fields from SAR, precipitation patterns from SMMR, and wind speed indications from these three instruments plus wind vectors from SASS. All measurements were in qualitative agreement. There were indications of precipitation-sized water drops causing errors in the integrated water vapor from SMMR. Consequently, the attenuation effect by cloud and rain water on the SASS winds could not always be resolved and corrected for. Little or no surface comparison data were available in these intense storms. Nevertheless, the value of simultaneous measurements from SMMR and SASS was clearly established. Comparisons to aircraft observations were carried out by Thompson et al. (1983). Their results suggest that SAR L-band (25 cm) observations in conjunction with scatterometer K-band (2.2 cm) observations provide better wind speed discrimination, in magnitude and spatial scale (tens of meters), than either instrument alone.

SASS was used by Hawkins and Black (1983) to verify advisories on the radius of gale-force winds produced in tropical cyclones. The scatterometer data revealed large errors of the standard radius in advisories. This was often due to asymmetry in the wind patterns around hurricanes and tropical cyclones, which makes the concept of a radius of strong winds unrepresentative. An overview by Black et al. (1985) of several tropical storms again demonstrated excellent agreement between Seasat, aircraft, and surface observations. Indeed, Seasat sensors opened a new era of measurements for hurricanes and typhoons. The practical products include: precipitation fields, wind vector and convergence/divergence fields, position of storm centers, radius or areas of gale-force winds, and initialization of models for storm surge wave prediction (Fig. 7).

6. Studies of ice

Because of the combination of microwave instruments on Seasat, unique intercomparison between signals over sea ice from SAR, SASS, SMMR, and the ALT

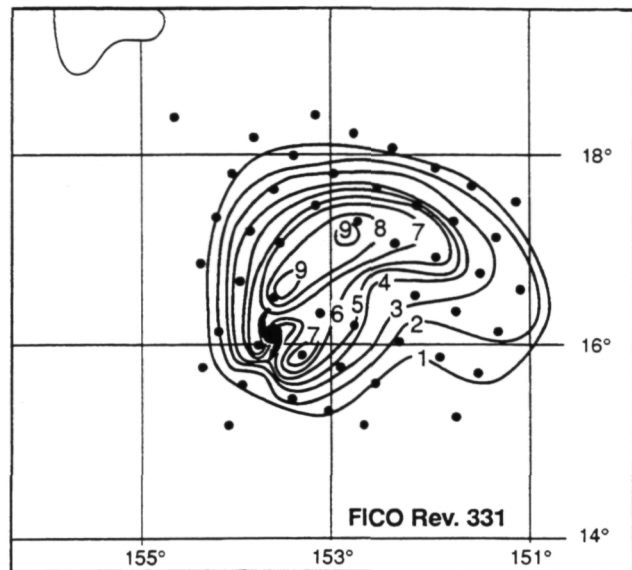


FIG. 7a. Rain rates in mm h^{-1} derived according to algorithm by Wentz, using 37- and 18-GHz Seasat SMMR data averaged to a 50-km grid for Hurricane Fico, on 0430 20 July 1978. The centers of the grid boxes are indicated by solid dots and the storm center is indicated by a hurricane symbol (after Black et al. 1985).

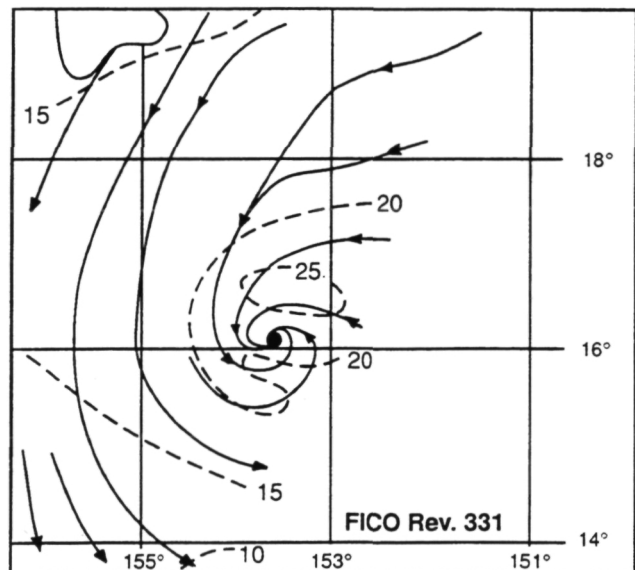


FIG. 7b. Isotach and streamline analyses for Hurricane Fico based on SASS wind aliases for 20 July 1978 that were closest in direction to the "surface truth" analysis (after Black et al. 1985).

were possible. SAR is evidently the most valuable instrument for sea ice studies, but data were only collected in selected portions of the orbit. Swift et al. (1985), presenting the conclusions from several case studies, claim that "the key cryospheric capabilities of the microwave instrument ensemble, which it was hoped to demonstrate with Seasat, have been accomplished." They included: identification of sea ice move-

ment, sea ice type and concentration, and position of the ice edge. Altimeter data have also provided maps of the surface waves and wind speeds in the marginal ice zones. All of the microwave sensors on Seasat were effective in locating the pack ice boundaries of the ocean. In addition, the VIRR proved useful in measuring the compactness of sea ice, and the SAR with its small footprint was used to plot ice motion by repeated views of the same floes (Hall and Rothrock 1981; Leberl et al. 1983; Carsey and Holt 1987). SAR, SASS, and SMMR can all be used to identify certain sea ice types as discussed by Carsey (1985), Holt and Digby (1985), and Carsey and Pikos (1990).

The variability of the volume and distribution of the masses of ice over the Antarctic continent and Greenland are of great interest when estimating changes in the global climate. Remy et al. (1989)

provide a database for global marine mass, heat, and momentum balances. Such data will be useful in initializing and updating numerical weather prediction models and improving the information in climate models.

In addition, new phenomena within the Seasat data are still being discovered. An example is the possibility to extract wind direction from SAR observations. An observation that SSM/I brightness temperatures show an upwind and downwind difference (F. Wentz 1990, personal communication) indicates that a microwave radiometer would be very useful for eliminating the 180° direction ambiguity that most modern designs of scatterometers still have. In turn, the detailed wind vector analyses from a scatterometer can be used to correct the radiometer-inferred SSTs and other algorithms for surface roughness effects on emissivity.

Such experiences gained from Seasat and subsequent satellite sensors may be useful for interpretation of sensors such as the future ERS-1 radar, which will be employed as both a synthetic aperture radar and a scatterometer. They

Eventually, the combination of satellite observations and numerical model results will allow calculations of turbulent surface fluxes and will provide a database for global marine mass, heat, and momentum balances. Such data will be useful in initializing and updating numerical weather prediction models and improving the information in climate models.

employed Seasat altimeter data to make detailed topographic maps of portions of the Antarctic ice sheet. They claim an accuracy of 0.4 m and expect to be able to observe any changes that may have occurred over a 12- to 13-year time span by employing altimeter data from the ERS-1 (Earth Resources Satellite-1), to be launched in 1991. Accuracy may be improved by applying corrections for ice metamorphosis due to katabatic winds, temperature, and other effects. These influences may be estimated by using microwave radiometer data synergistically (F. Remy 1990, personal communication).

7. Summary and discussion

The multiplicity of instruments on Seasat was valuable in many instances. Storm and ice studies have been mentioned. The combination of a microwave radiometer, a scatterometer, and sensors to estimate SSTs and PBL air temperatures will allow efficient global synoptic marine weather analysis from space. Surface waves, sea ice dynamics, and marginal ice zone characteristics are observable from space using microwave instruments. Eventually, the combination of satellite observations and numerical model results will allow calculations of turbulent surface fluxes and will

also speak strongly for simultaneous measurements from related sensors.

One element that was missing in the Seasat program was high-quality visible and infrared digital data in conjunction with the microwave data. Seasat did have the VIRR, Visible and Infrared Radiometer system, which unfortunately failed after 48 days of operation, i.e., during the hurricane and cyclone season. However, it had intentionally been chosen to be of low resolution to "match the resolution" of the microwave sensors (and probably also to keep the data rate low). It will be extremely valuable, especially for studying atmospheric storm systems with these new sensors, when more traditional satellite images of a quality comparable to the current NOAA AVHRR, Advanced Very High Resolution Radiometer, are obtained simultaneously. To the present, it has been very difficult to obtain collocated visible and infrared digital data of storms with any microwave data. Data from different satellites must usually be combined, and they have seldom been available in digital form post-facto and rarely coincident in time.

The sampling by Seasat's SMMR and SASS did not provide ideal overlap, neither for correcting SASS signals for atmospheric interference nor for combining SASS divergence fields with SMMR integrated water vapor or precipitation patterns. These examples of

omissions or mismatches constitute valuable lessons to be applied to a new era of instrument and sampling integration for enhanced benefits from satellite data.

We were motivated to present this review to the *Bulletin* audience because of the tremendous potential that passive and active microwave instruments in space have for improving meteorological analysis and prediction over the ocean. Awareness of this potential is needed because plans over the last decade to launch similar, follow-on instruments into space have often floundered. The reasons for this are perhaps economic but nonetheless hard to explain considering the success of the Seasat experiment. It is inevitable that instruments descendent in their design from the Seasat ones will eventually become operational. However, only the SMMR derivative, the Special Sensor Microwave/Imager (SSM/I), with five channels (three having almost identical frequencies to the atmospheric water sensing ones of SMMR) is in space. The first SSM/I was launched in 1987 on the operational F8 satellite of the Defense Meteorology Satellite Program and several more are scheduled. The Tropical Rainfall Measuring Mission (TRMM, Simpson et al. 1988) has recently received a "new start" status in NASA (J. Theon, personal communication). This satellite will travel in a low-altitude equatorial orbit to monitor tropical rain by radar and radiometer. The Geosat altimeter on its own satellite has provided good data for several years. It and the SSM/I have valiantly carried microwave remote sensing forward since Seasat days, but

One of the great losses to air-sea interaction studies has been the lack of a follow-on scatterometer to Seasat's SASS in the 1980s. The European Research Satellite, ERS-1, to be launched in June 1990, will have a single swath, three-antennae scatterometer. It is similar to the U.S. NSCAT, which is designed to eliminate one of the difficulties of SASS, namely the ambiguities in the wind direction (Wurtele et al. 1982; Levy and Brown 1986). However, it will furnish only a single scatterometer swath, and only a narrow swath from a single frequency microwave radiometer will be available along the satellite subtrack. The NSCAT, which is a two-swath, three-antennae scatterometer, is currently slated to be placed in orbit by the Japanese satellite, ADEOS, in 1995. A concurrent microwave radiometer on the same satellite is not yet a certainty.

We conclude that microwave radiometry and scatterometry can provide important services to weather forecasting and climate analysis, providing unprecedented data in midlatitudes, particularly in the Southern Hemisphere, in the tropics, and in polar regions as well. The accumulation of these new worldwide types of data on atmospheric properties should begin as soon as possible in order to provide the data-

base for climate evaluations and the data needed for projections of global climate changes. For instance, the differences between conventional analyses and scatterometer pressure fields suggest that climatological wind values for the Southern Hemisphere must be revised. With average wind speed from the Seasat SASS being 10% to 20% higher than those found in established climatologies, the implication is that flux estimates from bulk coefficient formulas will be at least that much higher for sensible heat, moisture, and carbon dioxide fluxes, and even greater for momentum flux.

We should now be carefully planning the optimal configurations of future systems with the diagnosing of atmospheric weather systems in mind. Such considerations are in progress within the program for an Earth Observing System (EOS), which aims to place many microwave sensors in orbit by the year 2000 on several satellite platforms. However, it is apparent that vigorous intellectual, as well as political, support from the meteorological community is essential for successful implementation of this ambitious and complicated satellite program.

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SATELLITE DERIVED INTEGRATED WATER VAPOR AND RAIN INTENSITY PATTERNS: INDICATORS OF RAPID CYCLOGENESIS

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1. INTRODUCTION

Rapidly intensifying extratropical cyclones are accompanied by severe weather conditions and frequently develop over the data sparse ocean areas. Although operational numerical model performance has improved in recent years, the strength can be poorly predicted, due in part to an inadequate initial analysis (Sanders, 1987). Recent research has demonstrated the importance of moist processes and latent heat release in the development of these cyclones (e.g. Kuo et al., 1991). Hence, an adequate representation of the moisture fields appears to be important for accurate prediction of these storms (Reed et al., 1988).

Atmospheric water vapor and precipitation intensity can be retrieved over the ocean from satellite microwave radiometers. These fields have been used to delineate frontal zones (McMurdie and Katsaros, 1985; Katsaros et al., 1989), to characterize typical water vapor distributions and content in storms occurring in various oceanic basins (McMurdie and Katsaros, 1991), and to study moisture pattern evolution in the Gulf of Mexico following the passage of a cold front (Rabin et al., 1991).

In this paper, we examine integrated water vapor fields and rain intensity patterns derived from the Scanning Multichannel Microwave Radiometer (SMMR) and Special Sensor Microwave/Imager (SSM/I) for several rapidly deepening and non-rapidly deepening midlatitude cyclones in the North Atlantic. Our goal is to identify features in the satellite data unique to the rapidly deepening cases, and to explore how these data can potentially be used in the analysis and forecasting of these events.

2. DATA ANALYSIS

The SMMR and the SSM/I are passive microwave instruments on the Nimbus 7 and the Defense Meteorological Satellite Program (DMSP) polar orbiting satellites, respectively. Information on each instrument is presented in table 1. The algorithms used to calculate SMMR integrated water vapor (IWV) are given in Chang et al. (1984) and Petty and Katsaros (1990) for the SSM/I. They are valid over ocean areas and are as accurate as radiosondes within a rms error of 2.1 and 2.0 kg m⁻², respectively. Since accurate measures of rainrate are difficult to obtain from satellite, we use a normalized polarization difference (P_{37}) in the 37 GHz channel to indicate rain intensity for the SSM/I cases. It is defined as the measured 37 GHz vertically polarized brightness temperature (T_b) minus the 37 GHz horizontally polarized T_b divided by the clear sky T_b polarization difference. Low values (0.1 or 0.2) indicate high rain intensity and high values (greater than 0.8) indicate no rain. For the SMMR cases, a simple polarization difference in the 37 GHz channel is used to indicate rain.

Thirty-four North Atlantic storms that occurred during the winters of 1979 - 1981 (SMMR) and 1988 (SSM/I) are divided into two groups: 1) Rapidly deepening

Table 1 - Comparison of SMMR and SSMI instruments

	SMMR	SSMI
Dates	1979 - 1985	1987 - present
Swath width	780 km	1394 km
Channels	6.6, 10.7, 18 GHz 21, 37 GHz	19, 22.2, 37, 85 GHz
Polarization	Vertical and Horizontal all channels	Vertical and Horizontal all channels, 22.2 in vertical only
Resolution (IWV, P_{37})	50 km x 50 km	50 km x 50 km

storms, where the storm experienced a deepening rate of 1.4 or greater for 24 hours (a deepening rate of one is defined in Sanders and Gyakum (1980) as a drop of 24 mb in 24 hours multiplied by $(\sin \phi / \sin 60^\circ)$ where ϕ is the latitude of the storm); and 2) Standard storms, where the storm experienced a deepening rate less than 1.4. The NMC hemispheric surface analysis is used to determine the central pressure and the position of the surface fronts. Water vapor fields and rainfall intensity are compared during the incipient (more than twelve hours prior to the period of most rapid deepening) and deepening stages.

IWV and P_{37} fields from individual passes over a storm are analyzed as follows. Only the passes that sampled the majority of the storm near the low center are included in the study. Maximum water vapor content in the warm sector, or along the cold and warm fronts, is estimated from the IWV patterns by picking the highest contour (contour interval is every 2 kg m⁻²) in the analyzed field along, or within 200 km of the frontal zones. In addition, the areal extent of the region of high water vapor content is estimated by counting the number of pixels greater than 30 kg m⁻², multiplying that number by the pixel resolution and rounding to the nearest 10³ km². The region considered does not extend south of 25°N or 5° of latitude south of the southern portion of the cold front in order to include just the warm sector and frontal zones sampled by the satellite. The value 30 kg m⁻² was chosen arbitrarily since it is approximately the average maximum IWV content for North Atlantic storms found by McMurdie and Katsaros (1991). For the rain intensity fields, the areal extent of the region of rain (quantified as P_{37} less than 0.8 for the SSM/I and 37 GHz polarization difference of less than 30°K for the SMMR cases) is calculated in the same manner. Maximum water vapor, areal extent of high vapor content and areal extent of precipitation from each pass over a storm are averaged together for rapidly deepening storms and standard storms. The averages in similar categories (e.g. maximum water vapor content during the incipient stage in rapid vs. standard cases) are tested to be significantly different using the two sample t-test.

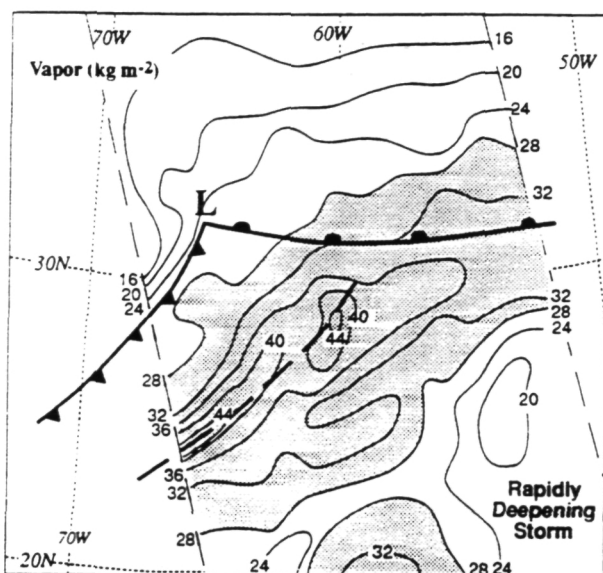


Figure 1 - I-WV from SSM/I for orbit 2970 at 10 UTC 16 January 1988. The field is contoured every 4 kg m^{-2} and the region greater than 28 kg m^{-2} is shaded. Frontal analysis is estimated from the NMC hemispheric surface charts at 06 and 12 UTC 16 January.

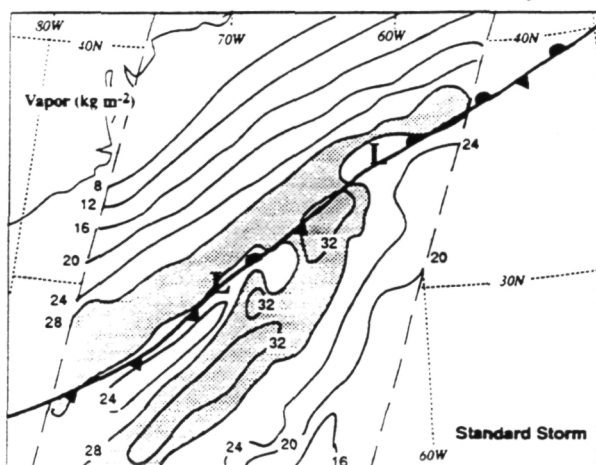


Figure 3 - I-WV from SSM/I for orbit 3274 at 23 UTC 6 February 1988. The field is contoured every 4 kg m^{-2} and the region greater than 28 kg m^{-2} is shaded. Frontal analysis is estimated from the NMC hemispheric surface charts at 00 UTC 7 February.

3. WATER VAPOR

3.1 Rapidly Deepening Case: 15 - 17 January 1988

The surface low of this system developed from a weak trough of 1018 mb situated at $32^\circ\text{N } 72^\circ\text{W}$ at 00 UTC 16 January. At this time a strong shortwave trough was situated along the east coast from Maryland southward. In six hours the surface low deepened to 1008 mb and a surface cold front is analyzed. By 12 UTC 16 January, an upper-level cut-off low formed with a height of 558 decameters situated just west of the now rapidly deepening surface low. The storm subsequently deepened to 992 mb and moved zonally to $32^\circ\text{N } 62^\circ\text{W}$ by 00 UTC 17 January. The storm achieved a maximum deepening rate of 1.5.

The I-WV field from the SSM/I at 10 UTC 16 January, prior to deepening, is shown in figure 1. In the region east of the cold front, along the prefront trough (dashed line in figure) and south of the warm front, the moisture content is high. Maximum values of 44 kg m^{-2} are evident along the trough. The region where the water vapor content is greater than 30 kg m^{-2} is $85.0 \times 10^4 \text{ km}^2$.

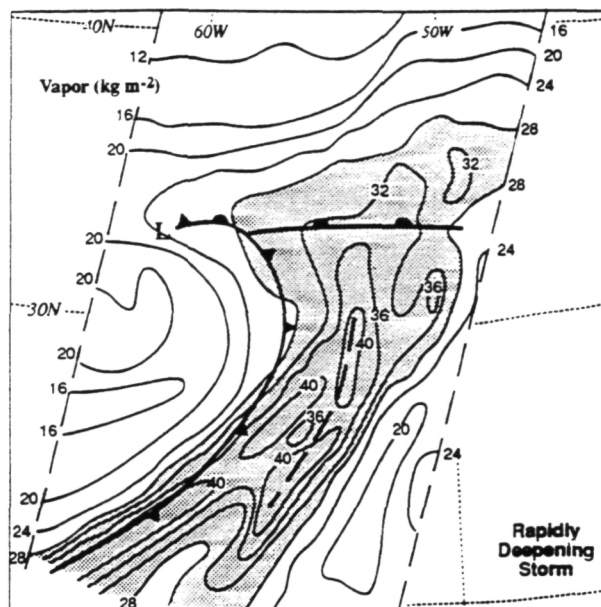


Figure 2 - I-WV from SSM/I for orbit 2977 at 22 UTC 16 January 1988. The field is contoured every 4 kg m^{-2} and the region greater than 28 kg m^{-2} is shaded. Frontal analysis is estimated from the NMC hemispheric surface charts at 18 UTC 16 Jan. and 00 UTC 17 Jan.

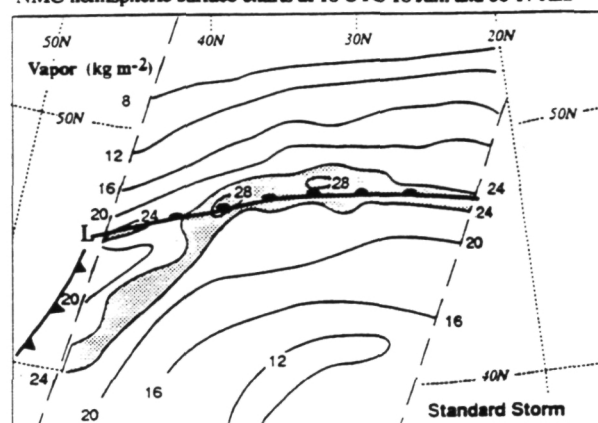


Figure 4 - I-WV from SSM/I for orbit 3287 at 21 UTC 7 February 1988. The field is contoured every 4 kg m^{-2} and the region greater than 24 kg m^{-2} is shaded. Frontal analysis is estimated from the NMC hemispheric surface charts at 18 UTC 7 February and 00 UTC 8 February.

In figure 2, the I-WV field at 22 UTC 16 January after rapid deepening is shown. The water vapor content is still high from the warm front southwards. Maximum water vapor content of 40 kg m^{-2} is evident at the trough and cold front, and the areal extent of water vapor content greater than 30 kg m^{-2} is still large, approximately $95.0 \times 10^4 \text{ km}^2$. The water vapor pattern has evolved into a comma shape, and the strength of the water vapor gradient behind the cold front has increased.

3.2 Standard Cyclone: 6 - 8 February 1988

This storm developed from a wave on a pre-existing surface cold front. At 12 UTC, a closed low of 1012 mb is analyzed at $30^\circ\text{N } 73^\circ\text{W}$ at the cold front. The surface low did not begin to deepen until 12 UTC 7 February when a rapidly moving upper-level shortwave trough moved just upstream of the surface low. The surface low progressed northeastward to $46^\circ\text{N } 54^\circ\text{W}$ and deepened to 992 mb by 00 UTC 8 February. The storm had a maximum deepening rate of 1.0.

The IWV field at 23 UTC 6 February, prior to deepening, is shown in figure 3. In contrast to figure 1, high water vapor content is confined to a narrow band along the front and in a narrow region south of the front. The maximum water vapor content in this region is only 32 kg m⁻², and the region where the IWV content is greater than 30 kg m⁻² is only 30.0 x 10⁴ km².

In figure 4, the IWV pattern at 21 UTC 7 February is shown. At this stage, the low has moved substantially northward and the water vapor content is significantly lower than in figure 3. The maximum water vapor content is 28 kg m⁻² along the warm front. The next orbit over this storm (Orbit 3288, not shown) samples the southern end of the cold front. Eventhough the front is south of 30°N, the maximum IWV content is also low, 36 kg m⁻², and confined to a narrow region along the front.

3.3 Summary of all cases

It was shown above that the rapidly deepening case exhibited substantially more water vapor content throughout its development than the standard case. In order to examine if this relationship holds for all cases, average maximum water vapor content and average area of IWV content greater than 30 kg m⁻² were calculated for all the storms examined to date and are displayed in tables 2 and 3 respectively. In table 2, the average maximum IWV content is higher in rapidly deepening storms than in standard storms during both stages. However, since the number of samples is small and the standard deviations large, the averages are not significantly different at a confidence level of 90% or greater.

In table 3, average area of IWV content greater than 30 kg m⁻² for storms is shown. Since the swath width of the SMMR is half that of the SSM/I, it is only possible to compare results from each instrument separately. In the SMMR standard cases, during the incipient stage, only three of the eight storms had any IWV values greater than 30 kg m⁻², and of those three, the areal extent was quite small. In contrast, the areal extent of high vapor content for the SMMR rapidly deepening cases during their incipient stage was on average a factor of ten times larger. Therefore, despite the very large standard deviations and small number of cases, the average area of IWV content greater than 30 kg m⁻² is significantly larger for SMMR rapidly deepening cases at the 95% confidence level. During the deepening stage, and for the SSM/I cases, the differences between rapidly deepening cyclones and standard cyclones are not significant due to large standard deviations and small number of cases.

Table 2 - Average maximum water vapor content for rapidly deepening and standard cyclones during the incipient and deepening stages. Ave denotes the average in kg m⁻², s.d. denotes the standard deviation and n the number of cases used.

A) RAPIDLY DEEPENING

	Incipient			Deepening		
	ave.	s.d.	n	ave.	s.d.	n
SMMR	27.6	7.3	6	32.5	6.5	12
SSM/I	36.7	7.0	3	36.2	4.7	4
together	30.4	8.1	9	34.2	6.1	16

B) STANDARD CYCLONES

	Incipient			Deepening		
	ave.	s.d.	n	ave.	s.d.	n
SMMR	26.4	5.5	9	31.3	5.0	13
SSM/I	34.3	2.3	2	35.0	1.4	2
together	27.9	5.9	11	31.7	4.8	15

Table 3 - Average area in warm sector where the IWV is greater than 30 kg m⁻² for rapidly deepening and standard cyclones. Ave denotes the average area in 10⁴ km², s.d is the standard deviation and n is the number of samples used.

A) RAPIDLY DEEPENING

	Incipient			Deepening		
	ave.	s.d.	n	ave.	s.d.	n
SMMR	53.6	78.5	5	35.6	20.1	8
SSM/I	43.6	13.1	3	56.2	32.2	4

B) STANDARD CYCLONES

	Incipient			Deepening		
	ave.	s.d.	n	ave.	s.d.	n
SMMR	5.5	9.1	8	27.3	23.6	13
SSM/I	34.3	21.2	2	30.1	1.2	2

4. RAIN

The rain distribution given by the P₃₇ fields from the SSM/I for the two storms discussed in the previous section are given in figures 5 and 6. The area covered by precipitation is quite extensive in the rapidly deepening case, figure 5. In addition, there are several areas of intense rainfall, evident as P₃₇ values of 0.2 or less located along the trough, cold front and north and west of the developing surface low. In order for the latent heat of condensation (by precipitation) to influence the development of the surface low, the laplacian of the heating, not just the heating itself, is important (e.g., see eq. 3 in Kuo et al., 1991). Near the low center and cold front, non-raining areas are in close proximity to raining areas. Although it is not possible to calculate the magnitude of the latent heating, the spatial pattern of precipitation alone indicates that the contribution to the development of the low by diabatic heating may be important in this case.

In figure 6, the precipitation area is smaller than in the rapidly deepening case. In addition, the regions of intense rainfall are much fewer and smaller in this case. Furthermore, the raining areas are more continuous in the vicinity of the developing low center and the spatial pattern of rain is less complex. Again, it is not possible to calculate the magnitude of the latent heating; however, the rain distribution indicates that the contribution by diabatic heat to the low development may be less in this case than in the rapidly deepening case.

In table 4, the average area covered by precipitation in rapidly deepening and standard cyclones are presented. Due to the different swath widths and the different ways of indicating rain, the SMMR and SSM/I cases must be considered separately. During the incipient stage, the SMMR standard cases had very light rainfall, and three of the eight cases exhibited no rain. The average rain area for incipient rapidly deepening cyclones is significantly larger than the average rain area for incipient standard cyclones at the 95% confidence level. In addition, the average rain areas are significantly different during the deepening stages at the 90% confidence level. The differences between rapidly deepening and standard cyclones for the SSM/I cases are not significant due to the small number of cases.

5. SUMMARY

In this paper we present IWV and P₃₇ analyses for thirty-four rapidly deepening and standard cyclones that occurred in the North Atlantic during the winters of 1979 - 1982 and 1988. We found that the maximum water vapor content was not significantly different in these two types of storms. Yet, during the incipient stage in the SMMR cases analyzed, the area of IWV amounts greater than 30 kg m⁻² in the warm sector and the area of precipitation were significantly larger in rapidly deepening cyclones than in standard cyclones. The examples presented in section 4

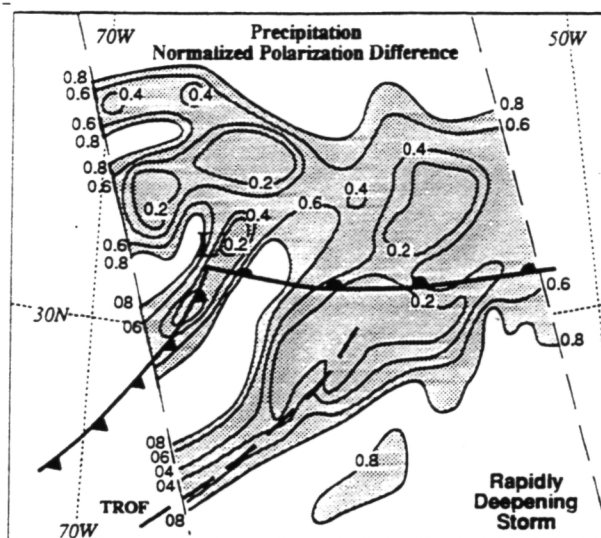


Figure 5 - P₃₇ from SSM/I for orbit 2970 at 10 UTC 16 January 1988. The field is contoured every 0.2 and the region less than 0.8 indicates rain and is shaded. Frontal analysis is estimated from the NMC hemispheric surface charts at 06 and 12 UTC 16 January.

Table 4 - Average rain area where P₃₇ is less than 0.8 for rapidly deepening and standard cyclones. Ave denotes the average area in 10⁴ km², s.d. is the standard deviation and n is the number of samples used.

A) RAPIDLY DEEPENING

	Incipient			Deepening		
	ave.	s.d.	n	ave.	s.d.	n
SMR	12.8	8.7	5	17.8	10.0	11
SSM/I	94.6	20.3	3	72.2	32.1	4

B) STANDARD CYCLONES

	Incipient			Deepening		
	ave.	s.d.	n	ave.	s.d.	n
SMR	3.5	3.7	8	10.4	11.5	10
SSM/I	128.1	74.9	2	99.8	67.9	2

demonstrated that differences in rainfall intensity and rainfall pattern variability may exist between rapidly deepening and standard storms. However, more cases need to be analyzed in order to quantify these differences. The results obtained from the cases studied to date are consistent with the hypothesis that moisture availability and latent heating are important in the development of rapidly deepening cyclones.

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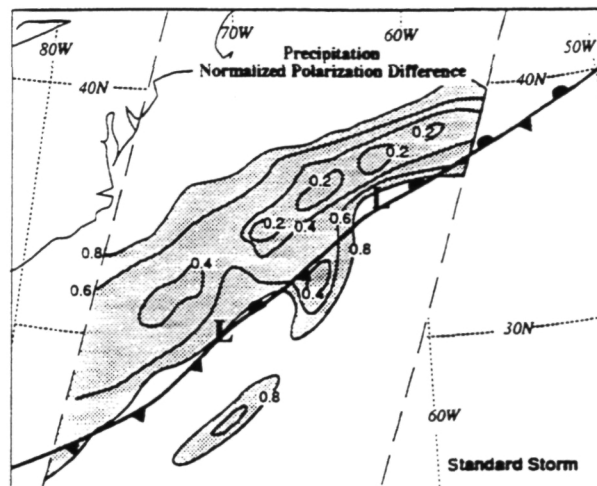


Figure 6 - P₃₇ from SSM/I for orbit 3274 at 23 UTC 6 February 1988. The field is contoured every 0.2 and the region less than 0.8 indicates rain and is shaded. Frontal analysis is estimated from the NMC hemispheric surface charts at 00 UTC 7 February.

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COMPARISON OF WIND SPEED MEASUREMENTS OVER THE OCEANS WITH THE SPECIAL SENSOR MICROWAVE/IMAGER AND THE GEOSAT ALTIMETER.

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1. INTRODUCTION

The Geosat radar altimeter, and the Special Sensor Microwave /Imager (SSM/I) have been used successfully to identify, locate, and analyze the evolution of atmospheric fronts and polar lows (Mognard and Katsaros, 1990; Claud et al., 1991). Using a combination of sequential passive and active microwave satellite sensors yields new insights on the location and the evolution of these atmospheric systems that cannot be obtained by more conventional (in-situ or visible and infra-red) instruments. These satellite microwave sensors will in the near future be used to provide input data for atmospheric and sea state numerical models. The Seasat satellite, that worked from June to October 1978, was the first satellite to have flown an ensemble of passive and active microwave instruments among which three: the altimeter, the scatterometer, and the Scanning Microwave Multichannel Radiometer (SMMR) gave estimates of the ocean wind speed. Comparison of the three wind speed estimates from Seasat showed that significant biases were present in the retrieved wind speed (Mognard and Campbell, 1984).

In order to compare wind speed estimates from the Geosat altimeter and the SSM/I, 25 colocated passes, within 2 hours of each other, have been selected and the SSM/I estimates of wind speed and atmospheric parameters extracted along the Geosat track. Both instruments and their algorithms are described in section 2. In section 3, a statistical comparison of wind speed estimates is presented and the effects of the atmospheric parameters derived from SSM/I, and of the sea state parameters from Geosat are analyzed. Quasi-simultaneous measurements by Geosat and SSM/I, along a Geosat track in the North-East Pacific, are presented in section 4.

2. THE INSTRUMENTS AND THEIR ALGORITHMS

2.1 The Special Sensor Microwave/Imager

SSM/I consists of seven separate radiometers which simultaneously measure the upwelling polarized microwave emission from the Earth and its atmosphere at four frequencies: 19.35, 22.2, 37, and 85.5 GHz. The spatial resolution ranges from 55 km at 19 GHz to 15 km at 85 GHz. A detailed description of the instrument is given by Hollinger et al. (1987).

The algorithms used in this paper for estimating atmospheric water vapor, liquid water, index of precipitation, and ice scattering index are by Petty and Katsaros (1990a). Estimates of integrated water vapor content (IWV) are obtained from a regression of brightness temperatures at 19 and 22 GHz. Integrated cloud liquid water content (ICLW) estimates use a combination of the 37 GHz vertically and horizontally polarized channels. The determination of the index of precipitation (P37) is based on the 37 GHz normalized polarization method of Petty and Katsaros (1990b) developed with the SMMR on Nimbus-7. As was the case with SMMR, P37 < 0.8 occurs only in the presence of precipitation, while P37 > 0.9 generally implies precipitation-free pixels. The P37 index of precipitation is used to identify all regions of rain, including stratiform and warm-cloud rain at 25 km resolution. An estimate of the concentration and size of frozen precipitation particles above the freezing level P85 is derived from a linear combination of brightness temperatures at 85 GHz and provides high resolution (15 km) estimates of the location and relative intensity of cold-cloud precipitation (Petty, 1990).

The Global D-Matrix algorithm of Goodberlet et al., 1989 (which uses a linear combination of the 19, 22, and 37 GHz channels) is used to retrieve the ocean surface wind speed in regions with no precipitation. Comparing SSM/I wind speed estimates to measurements from buoys in the National Data Buoy Center (NDBC) network, Goodberlet et al. (1989), obtained no bias and standard deviations ranging between 1.4 to 1.9 m/s with collocation within 25 km and 30 minutes.

The SSM/I wind speed estimates are compared to the Geosat radar altimeter measurements along the altimeter track. The atmospheric parameters deduced from the SSM/I are used to classify the data.

2.2 The Geosat radar altimeter

The Geosat altimeter is a short-pulse (3.125 ns) nadir-viewing radar operating at 13.5 GHz. The Geosat radar altimeter system is described by McArthur et al. (1987). The sea state estimates deduced from the radar altimeter and used in these analyses are the ocean backscatter coefficient derived from the automatic gain control loop used to normalize the amplitude of the ocean return signal, and the significant wave height

(SWH) computed on board the satellite from the slope of the mean return waveform edge. The Geosat sea state estimates are available every second or every 6.7 km along the satellite track.

The algorithm applied to the ocean backscatter coefficient to infer wind speed is described by Brown et al. (1981). In a 7-month comparison study with the NDBC data set, Dobson et al. (1987) found the Geosat altimeter wind speeds to have a 0.50 m/s low bias with a standard deviation of 1.6 m/s compared to the buoys.

3. STATISTICAL COMPARISON

To compare SSM/I and Geosat wind speed estimates, a data set of collocated measurements has been extracted. This data set is composed of satellite passes that occurred over the world oceans during the first week of August 1987 and have a time difference less than 2 hours. From the 1394-km wide swath, only the SSM/I measurements collocated along the 10-km wide Geosat tracks have been selected. In order to compare the two wind speed estimates over similar size footprint, the altimeter data have been averaged over 4 s which approximately corresponds to a footprint of 10x30 km. Using trial and error, a region around the center of the altimeter footprint of $0.2^\circ \times 0.2^\circ$ in latitude and longitude has been selected as the optimum area over which to average the SSM/I data for comparison with the altimeter wind speed. A total of 25 Geosat and SSM/I passes have thus been selected and a scatter diagram of the wind speed estimates by the two instruments is presented in Figure 1 where 1616 measurements are compared. There is essentially no bias between the two data sets, on the average the altimeter measures winds that are 0.13 m/s higher than the SSM/I, and the standard deviation of the wind speed difference is 1.72 m/s which is of the same order of magnitude as the comparison of each sensor with the NDBC buoy network (Dobson et al., 1987; Goodberlet et al., 1989).

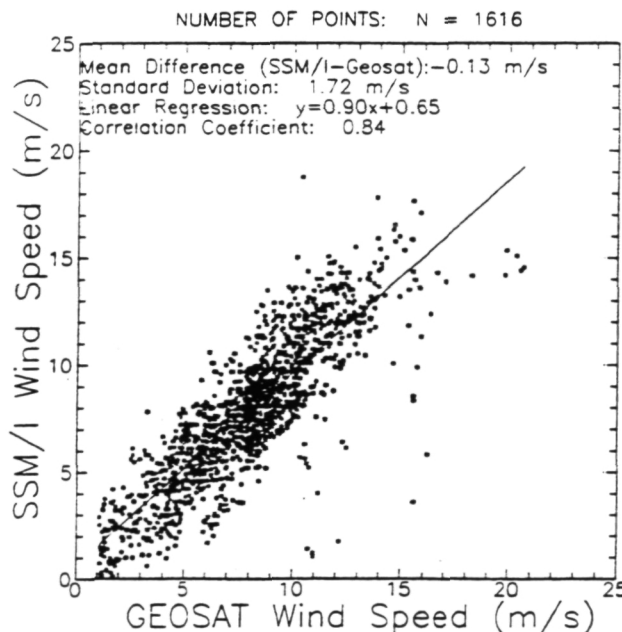


Figure 1: Scatter diagram of SSM/I versus Geosat altimeter wind speed estimates for the global collocated data set with a maximum time interval of 2 hours.

The SSM/I data set consists of wind speed as well as atmospheric parameters estimate (see section 2.1): integrated water vapor (IWV), integrated cloud liquid water (ICLW), index of precipitation (P37), and the ice scattering index (P85). The SSM/I wind speed estimation is only valid in regions with no rain (Goodberlet et al., 1989). According to the values of the precipitation index ($P37 < 0.8$), there are only 7 data points in the original data set that are contaminated by rain and should have been eliminated (Table 1). Statistical comparisons as a function of the different atmospheric parameters value derived with SSM/I, show a significant increase in the degree of confidence for data points associated with small amount of integrated cloud liquid water, less than 0.1 kg/m^2 (Table 1). In cases when $ICLW < 0.1 \text{ kg/m}^2$, the standard deviation of the mean wind speed difference decreases to 1.47 m/s. In intermediate cases of $ICLW > 0.1 \text{ kg/m}^2$ but no precipitation the standard deviation is 1.62 m/s (Table 1).

TABLE 1: Wind speed statistical comparison for collocated Geosat and SSM/I passes with time differences less than 2 hours. N is the number of data points. SSM/I minus Geosat wind speed and the corresponding standard deviation, in parenthesis, are listed in the column labelled SSM/I - Geosat with the correlation coefficient in column CC.

	N	SSM/I-Geosat	CC
Global data set	1616	-0.13 (1.72)	0.84 (1)
No Rain	1609	-0.15 (1.62)	0.86 (2)
ICLW < 0.1 kg/m^2	1130	-0.04 (1.47)	0.88 (3)

To check the influence of the time difference between the two satellite passes on the statistical parameters, a new data set with time differences less than one hour has been extracted. The statistics obtained with this new data set are presented Table 2 and Figures 2 and 3. Eighteen passes are now considered with 1056 comparison points where the precipitation index P37 is higher than 0.8 (no rain contamination) (Table 2, line (a)). For data points with less than one hour difference, the standard deviation of the differences decreases from 1.62 m/s (data set with a maximum of 2 hours difference, (Table 1, line (2)) to 1.42 m/s (Table 2, line (a)), and Figure 2), and from 1.47 m/s (Table 1, line (3)) to 1.18 m/s for pixels with small amount of cloud liquid water ($ICLW < 0.1 \text{ kg/m}^2$), (Table 2, line (b)), and Figure 3). The decrease in standard deviation shows that part of the discrepancies is due to time differences larger than one hour.

Possible effect of wave height on wind speed differences between the two instruments can be studied using the significant wave height (SWH) measurement by the altimeter. Statistics obtained for three subsets of wave heights are also presented Table 2, lines (a.1) to (a.3). The discrepancy in the statistics between the data sets with waves less than 4 m (Table 2, lines (a.1), (a.2)) and waves higher than 4 m (Table 2, line (a.3)) disappears in the data sets with $ICLW < 0.1 \text{ kg/m}^2$ (Table 2, lines (b.1), (b.2), and (b.3)). When there is only small amounts of cloud liquid water in the atmosphere ($ICLW < 0.1 \text{ kg/m}^2$, Table 2, lines (b)), there are no significant differences in the statistical parameters as a function of SWH. A significant difference in the statistical parameters arises when $SWH > 4 \text{ m}$ are associated with amounts of cloud liquid water larger than 0.1 kg/m^2 which correspond to the crossing of atmospheric fronts, and stormy regions.

TABLE 2: Wind speed statistical comparison for collocated Geosat and SSM/I passes with time differences less than 1 hour. The columns are the same as in Table 1. The statistical comparison is obtained as a function of SWH measured by the altimeter.

	N	SSM/I-Geosat	CC	
No Rain	1056	-0.46 (1.42)	0.88	(a)
0<SWH<2m	463	-0.38 (1.19)	0.84	(a.1)
2<SWH<4m	430	-0.56 (1.26)	0.87	(a.2)
4<SWH<6m	163	-0.41 (2.18)	0.67	(a.3)
ICLW<0.1kg/m ²	639	-0.44 (1.18)	0.91	(b)
0<SWH<2m	281	-0.35 (1.09)	0.88	(b.1)
2<SWH<4m	266	-0.58 (1.22)	0.88	(b.2)
4<SWH<6m	92	-0.29 (1.26)	0.84	(b.3)

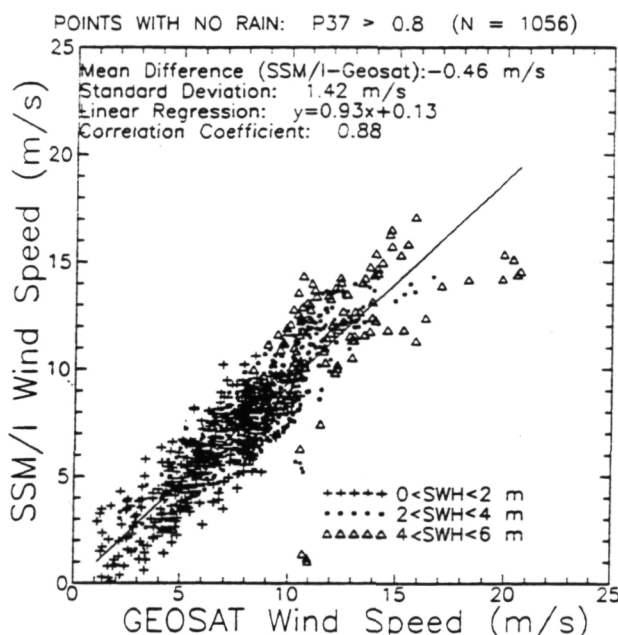


Figure 2: Scatter diagram of SSM/I versus Geosat altimeter wind speed estimates for a maximum time interval of 1 hour. Different symbols display different SWH values derived from the altimeter measurement.

4. EXAMPLE OF QUASI-SIMULTANEOUS MEASUREMENTS BY GEOSAT AND SSM/I

A pass in the North-East Pacific on August 5, 1987, has been chosen to illustrate the quasi-simultaneous measurement by Geosat and SSM/I along the altimeter track. There is only 12 minutes difference between the two satellite passes (Geosat at 0412 UTC and SSM/I at 0424 UTC) and less than 2 hours time difference from the corresponding weather map for 0600 UTC where the Geosat ascending track has been superimposed (Figure 4). The altimeter track starts in a region of relatively moderate winds (ship located at approximately 150°W and 41°N reports 20 kn) then crosses a col with low winds (ship reports 10 kn at 152°W and 45°N) before crossing a cold front at 156°W and 49°N and reaching a region of strong atmospheric pressure gradient behind the front where ships report wind speed between 20 and 35 kn.

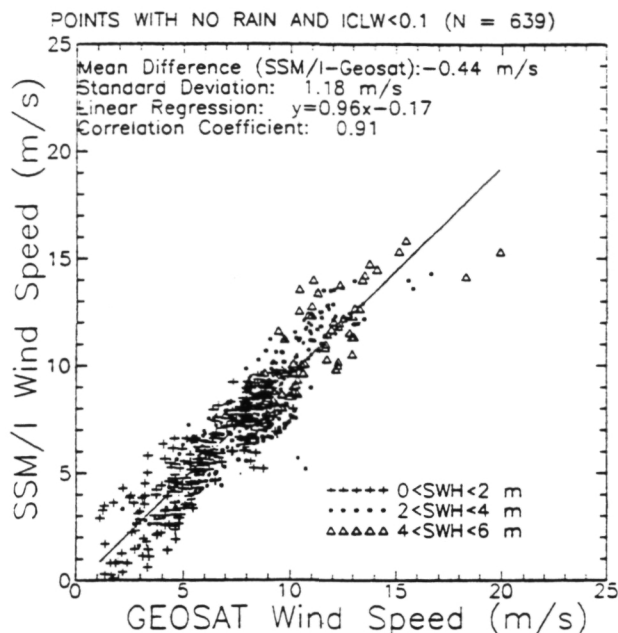


Figure 3: Scatter diagram of SSM/I versus Geosat altimeter wind speed estimates for pixels with small amount of cloud liquid water.

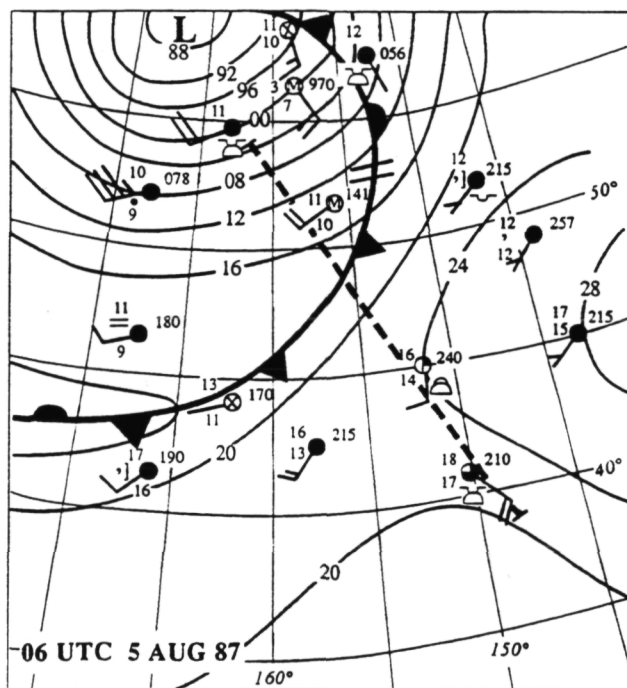


Figure 4: Synoptic weather map with ship reports and Geosat track superimposed (dashed line).

4.1 The SSM/I atmospheric parameters

Figure 5 presents the atmospheric parameter variations derived from SSM/I along the Geosat track. The integrated water vapor variations show the characteristic water vapor maximum in the warm air sector ahead of the cold front, accompanied by a sharp decrease in the cold air mass beyond the front (Figure 5a). The strong water vapor gradient (Figure 5a) located at approximately 48°N is a characteristic feature of the cold front crossing shown on the weather map (Figure 4). The location of SSM/I water vapor gradients

exceeding $0.1 \text{ (kg/m}^2\text{)/km}$ provides an excellent objective indication of the location of many surface cold fronts (Katsaros et al., 1989) as is the case here. The precipitation index (P37), which may be interpreted as a measure of the visibility of the polarized sea surface through cloud and precipitation (Petty, 1990) exhibits a similar feature as the IWV with a strong gradient located at about 49°N (Figure 5b). This gradient corresponds to a region with $P37 < 0.8$ indicating heavy precipitations during the crossing of the frontal zone. Another region of heavy rain is behind the front between 52 and 54°N . The calm region of the col south of the front (Figure 4) from about 44 to 47°N corresponds to precipitation-free pixels with P37 of the order of 0.95 . P37 values between 0.8 and 0.9 , imply at most light precipitation. The vertically integrated cloud liquid water variations along the Geosat track are presented in Figure 5c. In the region of the front crossing, the ICLW have been flagged because of heavy precipitation. South of the front, the col is the region with no precipitation ($P37 < 0.9$) and little cloud liquid water ($\text{ICLW} < 0.1 \text{ kg/m}^2$). The region of precipitation ($P37 < 0.8$) north of the front corresponds to ICLW between 0.28 and 0.34 kg/m^2 . Everywhere else, the values of ICLW vary between 0.1 and 0.3 kg/m^2 , an indication of diverse cloud structures

4.2 SSM/I and altimeter wind speed comparison

SSM/I and Geosat both measure sea surface wind speed. The wind speed and significant wave height variations measured by the altimeter are presented in Figure 6. At the start of the pass (southward), the altimeter measures 2 m waves and 9 m/s wind speed in agreement with the 20 kn ship report in the vicinity. The crossing of the col is signaled by very low wave (about 1 m) and by low winds varying between 1 and 5 m/s . In the North, the crossing of the cold front is marked by a wind speed increase from 5 to 10 m/s and a region where waves increase to 2 m . Gradients of wind speed and wave height are measured to the north, beyond the cold front, across the atmospheric pressure gradient zone. In section 4.1, the portions of the pass affected by heavy rains (across the front and in part of the region of strong atmospheric pressure gradient) where the SSM/I wind speed cannot be derived are discussed. In the portions with no heavy precipitation, an estimate of the SSM/I wind speed is obtained and plotted in Figure 7 along with the Geosat estimates. A very good global agreement is obtained especially in regions of wind gradients. Only in the southern portion of the pass does SSM/I measure slightly higher winds than Geosat. Part of these discrepancies can be accounted for by the differences in footprint size between the two sensors.

5. CONCLUSION

A comparison of colocated wind speed measurements by the Geosat altimeter and the SSM/I within a 2-hour time frame yields a statistical agreement of the same order of magnitude as was obtained for the comparison of the individual satellite sensors with the NDBC buoy network (Dobson et al., 1987; Goodberlet et al., 1989; Figure 1; Table 1). There is no significant bias between altimeter and SSM/I wind speed and a standard deviation of 1.62 m/s after complete elimination of rain pixels (Table 1, line(2)). When the time frame for the comparison is reduced to 1 hour, the number of data points decreases by about one-third, the altimeter has a 0.5 m/s high bias compared to the SSM/I and the standard deviation decreases to 1.42 m/s for the global data set and to 1.18 m/s for conditions when there is little cloud liquid water ($\text{ICLW} < 0.1 \text{ kg/m}^2$). The agreement between the two satellite sensors under favorable atmospheric conditions is much better than the

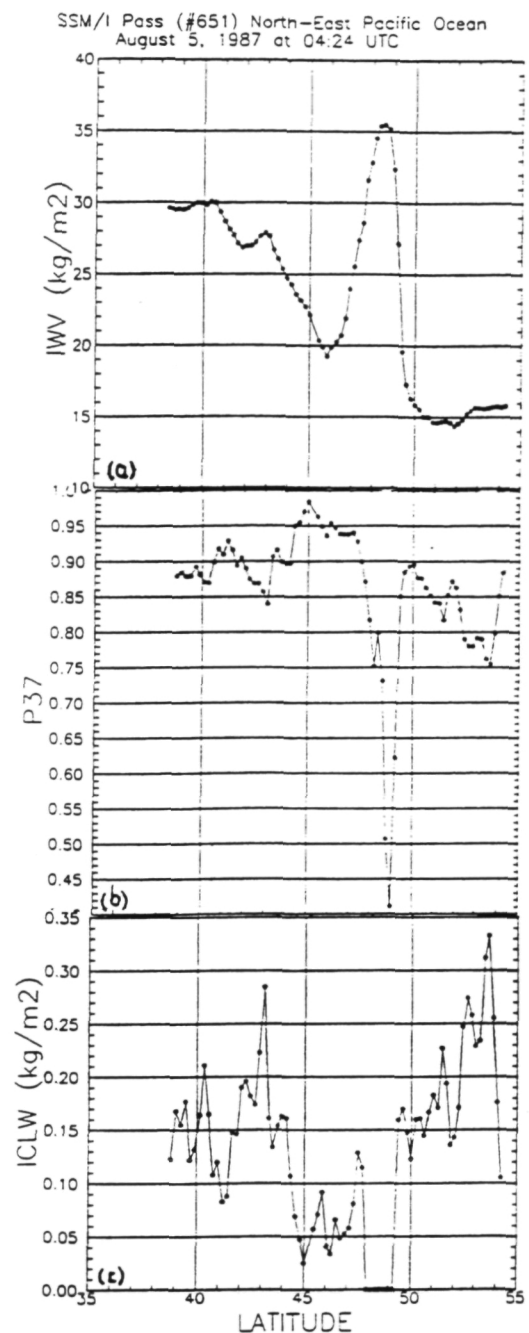


Figure 5: Variations of atmospheric parameter estimates from SSM/I along the Geosat track; (a) integrated water vapor variations; (b) precipitation index; (c) integrated cloud liquid water (pixels where ICLW could not be estimated are arbitrarily set to zero).

agreement of the individual instruments with the buoy network. This should not be too surprising since both Geosat and SSM/I measure spatially averaged wind speeds over regions of $30\text{--}50 \text{ km}$ size over a time interval of a few seconds while buoys measure time averaged winds (over several minutes) in one point location. The effect of wave height on the statistics is negligible away from storms ($\text{ICLW} < 0.1 \text{ kg/m}^2$) in regions where sea state is swell dominated (Table 2). In the vicinity of storms atmospheric and maybe sea state effects are responsible for the poor correlation between the two wind speed measurements and for the standard deviation increase (Table 2, line (a.3)).

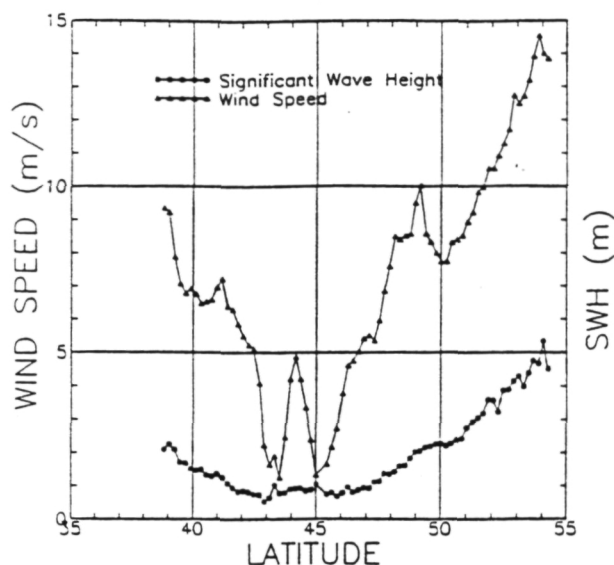


Figure 6: Geosat radar altimeter wind speed and significant wave height (SWH) estimates.

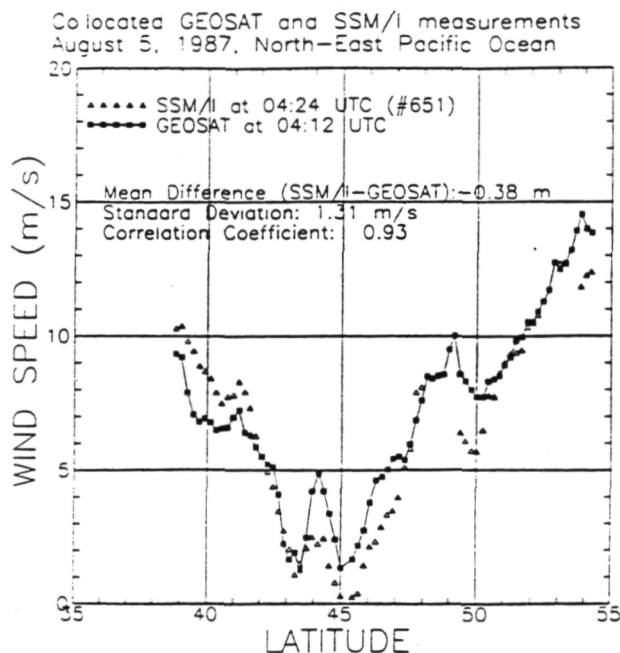


Figure 7: Geosat and SSM/I wind speed variations at 19.5 m above sea level.

The quasi-simultaneous Geosat and SSM/I measurements across a region of low winds, then across a cold front and in the region of strong atmospheric pressure gradient behind the front demonstrate the ability of both instruments to accurately locate and measure atmospheric and sea state gradients. In regions with little or no precipitation, the two wind speed estimates have a correlation coefficient of 93%. Compared to buoys or ship reports of wind, the satellite winds have better or comparable accuracy and are the only instruments able to provide ocean wide gradients (Geosat and SSM/I) and fields (SSM/I) of atmospheric or sea state parameters. These satellite measurements should be used in combination with in situ measurements and models to improve our understanding of the interactions between the atmosphere and the ocean.

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MORNING-EVENING DIFFERENCES IN GLOBAL AND REGIONAL
OCEANIC PRECIPITATION AS OBSERVED BY THE SSM/I

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1. INTRODUCTION

Among the important unanswered scientific questions regarding global precipitation are the following: (1) What is the diurnal cycle of tropical oceanic rainfall and how does it vary in space? and (2) what are the relative contributions of convective and stratiform precipitation, and how does their ratio vary in different regions and in different seasons?

While a definitive answer to these and similar questions may have to await the successful conclusion of the Tropical Rainfall Measuring Mission (TRMM) in the late 1990s, we believe relevant information can already be extracted from the data of an operational satellite sensor, the Special Sensor Microwave/Imager (SSM/I). This view is supported by the following preliminary global SSM/I statistics, which appear to show significant morning-evening differences both in the prevalence of moderate to heavy rain and in the mean intensity of convection within those areas unambiguously identified as raining.

Over the open ocean, meaningful observations of the global-scale spatial distribution and temporal evolution of tropical precipitation are currently feasible from satellites alone. To date, much of what is known about the diurnal cycle of convection located well away from land masses is based on infrared observations of cloud tops. For example, Albright et al. (1985) used the fractional coverage by cloud tops with temperatures colder than -36°C , as observed by the GOES-West geostationary satellite, to demonstrate a pronounced diurnal cycle in deep convection over the Central Tropical Pacific. In the Intertropical Convergence Zone (ITCZ), this cycle was found to have a distinct morning maximum and evening minimum, while in other areas, such as the South Pacific Convergence Zone (SPCZ), an afternoon maximum was observed.

In another study, Hartmann and Recker (1986) used nine years of data from the polar-orbiting NOAA series of satellites to compute the diurnal harmonic in outgoing longwave radiation (OLR) throughout the tropical belt. Because diurnal variations in low-level and high-level clouds tended to be 180° out of phase in regions of intense convection over the ocean, the amplitude of the diurnal variation in OLR was perceived to be "a weak reflection of more substantial variations in cloud type in this area."

All such studies to date, excluding those relying on surface observations from a few widely spaced islands or on data obtained during tropical mesoscale field experiments (e.g., Gray and Jacobson 1977, and studies cited therein), have provided only an indirect measure of the diurnal variations in precipitation associated with tropical oceanic convection. Infrared and visible satellite sensors observe only cloud top parameters, which in the tropics implies mainly the cirrus anvils and blow-off of convective towers.

Satellite microwave techniques are known to offer a more physically direct indication of precipitation activity, and because of the selection of channels available on current microwave sensors like the SSM/I, various 'modes' of interpreting microwave data (e.g., scattering-based vs emission- or attenuation-based algorithms) are available which can potentially be used to extract independent information concerning certain aspects of a storm's microphysical structure and overall intensity. Microwave brightness temperatures are generally unaffected by the non-precipitating cirrus shields which pervade the tropics and play havoc with infrared remote sensing techniques.

Table 1: Definitions of geographic regions.

Region	Latitude	Longitude
1 Tropics	$25^{\circ}\text{ S} - 25^{\circ}\text{ N}$	—
2 TWP	$25^{\circ}\text{ S} - 25^{\circ}\text{ N}$	$120^{\circ}\text{ E} - 170^{\circ}\text{ W}$
3 TCP	$25^{\circ}\text{ S} - 25^{\circ}\text{ N}$	$170^{\circ}\text{ W} - 130^{\circ}\text{ W}$
4 TEP	$25^{\circ}\text{ S} - 25^{\circ}\text{ N}$	$130^{\circ}\text{ W} - 80^{\circ}\text{ W}$
5 ITCZ	$0^{\circ} - 15^{\circ}\text{ N}$	$170^{\circ}\text{ W} - 130^{\circ}\text{ W}$
6 SPCZ	$25^{\circ}\text{ S} - 0^{\circ}$	$170^{\circ}\text{ W} - 130^{\circ}\text{ W}$
7 TIO	$25^{\circ}\text{ S} - 25^{\circ}\text{ N}$	$40^{\circ}\text{ E} - 100^{\circ}\text{ E}$
8 GATE	$5^{\circ}\text{ N} - 15^{\circ}\text{ N}$	$30^{\circ}\text{ W} - 20^{\circ}\text{ W}$
9 NML	$30^{\circ}\text{ N} - 60^{\circ}\text{ N}$	—
10 SML	$60^{\circ}\text{ S} - 30^{\circ}\text{ S}$	—

2. DATA AND METHODOLOGY

The SSM/I carries 19.35, 37.0, and 85.5 GHz channels in dual polarization (22.235 GHz is available only in vertical polarization). The dual polarization brightness temperatures at 19.35 and 37.0 GHz lend themselves to adaptations of the Petty and Katsaros (1990, 1991) attenuation-based algorithm, in which the polarization difference is interpreted as a measure of the visibility of the sea surface and thus of effective rain cloud opacity. Alternatively, using the 37.0 and 85.5 GHz channels, one may seek to isolate the so-called scattering signal (e.g., Spencer 1986, Spencer et al. 1989) associated with moderate to heavy cold-cloud precipitation containing precipitation-size ice particles above the freezing level.

While the twice-daily, non-overlapping sampling of the current SSM/I is not ideal for estimating total daily rainfall nor for studying the entire diurnal cycle of precipitation, its sun-synchronous morning-evening orbit coincides well with the maxima and minima of convective activity observed or inferred over many areas of the tropical ocean.

For the present preliminary analysis of oceanic rainfall statistics, global oceanic SSM/I data were simply scanned for pixels which exhibited a 37 GHz polarization difference (vertically polarized brightness temperature minus horizontally polarized brightness temperature) of less than 15 K. Such a low polarization difference over the open ocean is a completely unambiguous indication of moderate to intense precipitation. Co-located brightness temperatures from all seven channels of the SSM/I were saved for each pixel so identified. Bad scans and geographically mislocated blocks of data were objectively identified and removed from the resulting data base.

Regions of lighter rainfall or isolated convective cells are likely to escape detection using the threshold approach described above. Indeed, the data of Petty and Katsaros (1991) suggest that, in the subtropics at least, the chosen polarization threshold of 15 K will cap-

Table 2: Number of SSM/I pixels with 37 GHz polarization differences of less than 15 K, for the period 20 July 1987 — 19 August 1987 (by latitude belt).

Latitude	AM	PM	AM/PM
45°N – 60°N	3,351	2,737	1.22
30°N – 45°N	7,737	4,764	1.62
15°N – 30°N	4,078	3,265	1.25
0° – 15°N	20,927	16,313	1.28
15°S – 0°	6,359	4,463	1.42
30°S – 15°S	6,322	5,054	1.25
45°S – 30°S	7,877	8,037	0.98
60°S – 45°S	769	640	1.20
Total	57,420	45,273	1.27

Table 3: Number of SSM/I pixels with 37 GHz polarization differences of less than 15 K, for the period 20 July 1987 — 19 August 1987 (by region).

Region	AM	PM	AM/PM
1 Tropics	33,971	25,895	1.31
2 TWP	5,978	5,327	1.12
3 TCP	7,428	6,525	1.14
4 TEP	6,524	5,649	1.15
5 ITCZ	3,606	3,094	1.17
6 SPCZ	3,681	3,389	1.09
7 TIO	7,456	4,352	1.71
8 GATE	772	732	1.05
9 NML	11,085	7,501	1.48
10 SML	8,646	8,677	1.00

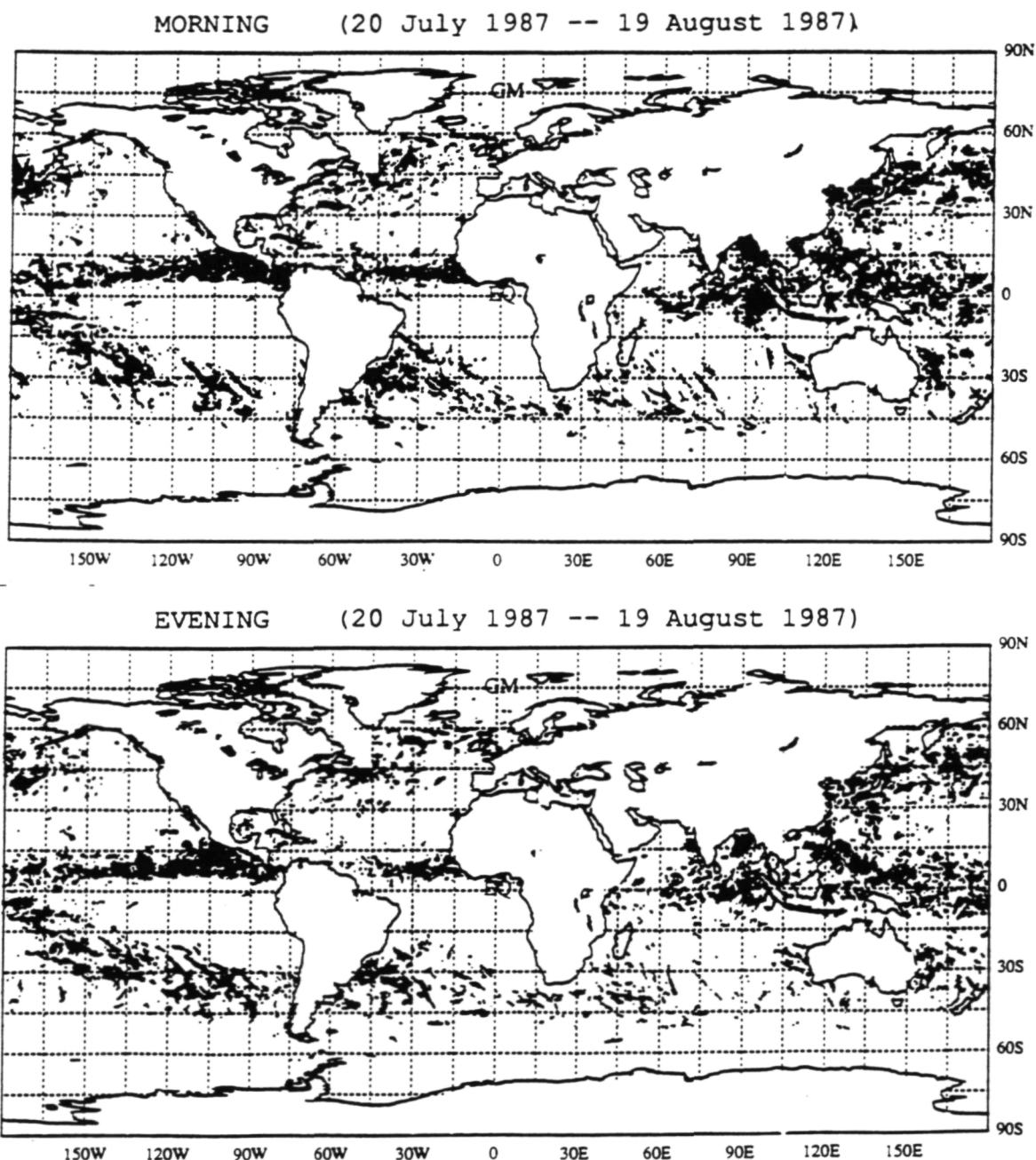


Fig. 1 Locations of SSM/I pixels with 37 GHz polarization differences of less than 15 K, for the period 20 July 1987 — 19 August 1987.

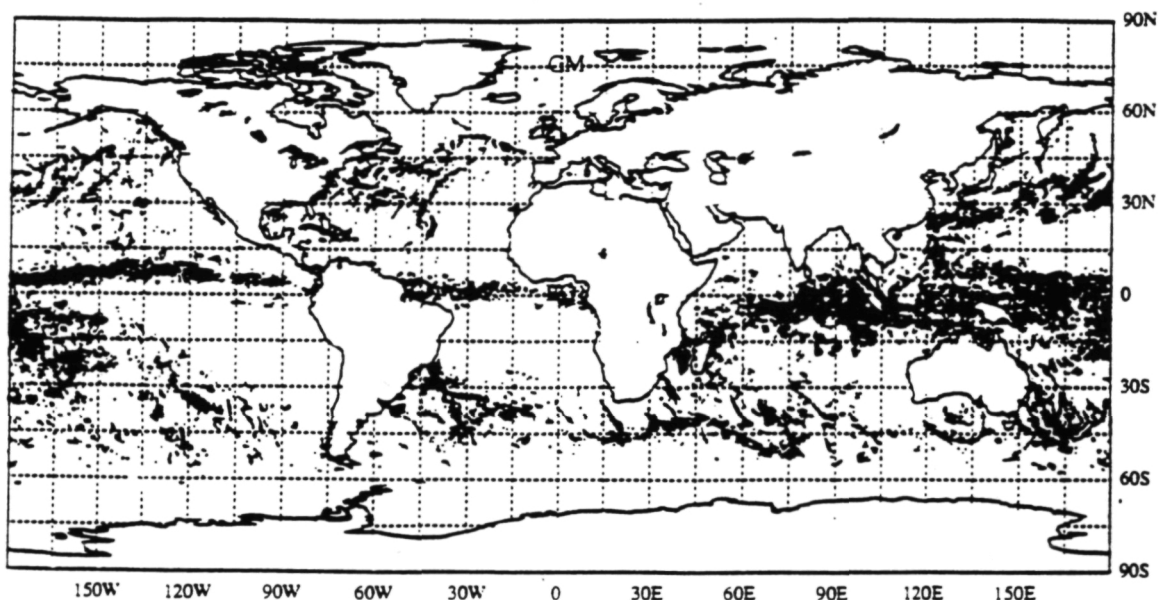
Table 4: Number of SSM/I pixels with 37 GHz polarization differences of less than 15 K, for the period 13 January 1988 — 12 February 1988 (by latitude belt).

Latitude	AM	PM	AM/PM
45°N - 60°N	2,067	2,251	0.92
30°N - 45°N	6,576	6,316	1.04
15°N - 30°N	4,108	3,530	1.16
0° - 15°N	14,476	9,124	1.59
15°S - 0°	16,760	9,810	1.71
30°S - 15°S	7,610	6,646	1.15
45°S - 30°S	10,040	10,161	0.99
60°S - 45°S	3,711	3,782	0.98
Total	65,348	51,620	1.27

Table 5: Number of SSM/I pixels with 37 GHz polarization differences of less than 15 K, for the period 13 January 1988 — 12 February 1988 (by region).

Region	AM	PM	AM/PM
1 Tropics	38,546	24,536	1.57
2 TWP	9,609	4,572	2.10
3 TCP	12,507	7,635	1.64
4 TEP	1,856	1,743	1.06
5 ITCZ	5,332	2,764	1.93
6 SPCZ	6,921	4,666	1.48
7 TIO	9,590	7,005	1.37
8 GATE	0	0	—
9 NML	8,639	8,561	1.01
10 SML	13,751	13,943	0.99

MORNING (13 January 1988 -- 12 February 1988)



EVENING (13 January 1988 -- 12 February 1988)

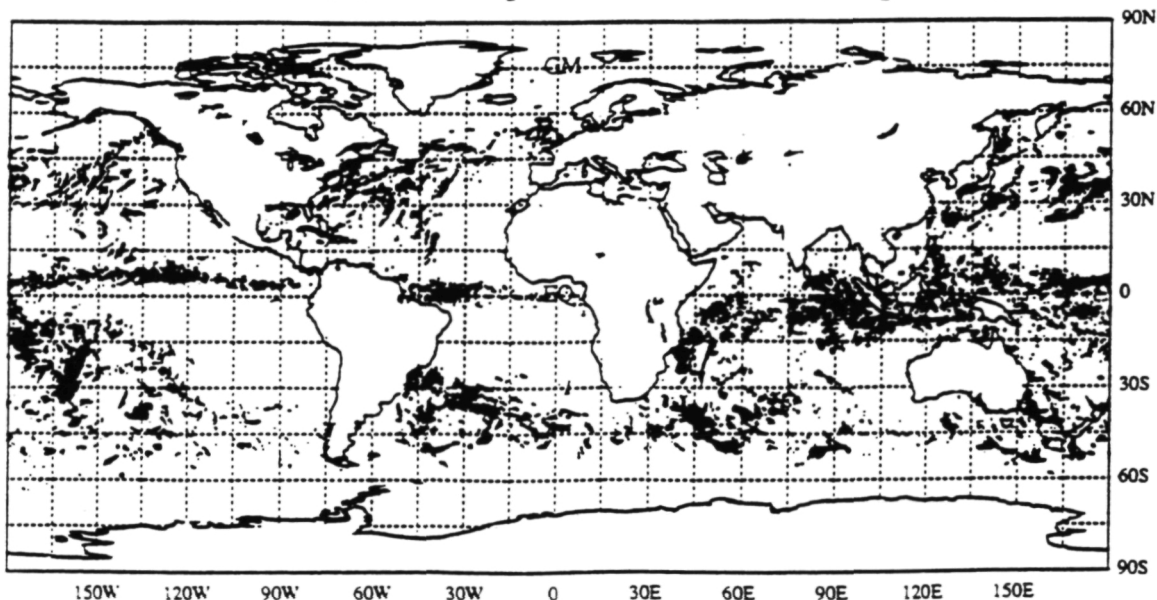


Fig. 2 Locations of SSM/I pixels with 37 GHz polarization differences of less than 15 K, for the period 13 January 1988 — 12 February 1988.

ture only about one fourth (by volume) of the total rainfall sampled by the SSM/I. Moreover, since the polarization threshold was not adjusted to account for regional variations in water vapor amount, surface roughness, and rain layer depth, the percentage of rainfall which escapes detection will vary from one storm to the next and, in particular, from the tropics to the midlatitudes. However, since our initial objective was simply to establish whether significant morning-evening differences could be discerned in the SSM/I data, irrespective of the absolute value of the total rainfall, the obvious limitations of a fixed threshold method can be tolerated.

We collected global oceanic rainfall data for two time periods, each one month in length. The first period (20 July–19 August 1987) coincides with the peak of the Northern Hemisphere summer. Because it also falls in the middle of the June–September season during which the GARP Atlantic Experiment (GATE) was conducted in 1974, results obtained from the SSM/I for the GATE area may be reasonably compared with the earlier GATE results. The second period (13 January–12 February 1988) coincides with the Northern Hemisphere winter, and SSM/I results for the central tropical Pacific may be compared with those of Albright et al. (1985) obtained for the period January–February 1979.

For each of the two periods, data were classified according to whether they corresponded to a morning pass (approximately 0600 LST at the equator; Figs. 1a and 2a) or an evening pass (approximately 1800 LST; Figs. 1b and 2b). The data were further stratified by 15° latitude belts and by specific geographic regions. Regions considered include (1) the Tropics, (2) the Tropical Western Pacific (TWP), (3) the Tropical Central Pacific (TCP), (4) the Tropical Eastern Pacific (TEP), (5) a portion of the Pacific Intertropical Convergence Zone (ITCZ), (6) the Southern Pacific Convergence Zone (SPCZ), (7) the Tropical Indian Ocean (TIO), (8) the GATE region, (9) Northern Midlatitudes (NML), and (10) Southern Midlatitudes (SML). Definitions of these regions are given in Table 1.

3. RESULTS

A simple count of the number of saved pixels gives a consistent and clear picture of significant morning-evening differences in the areal coverage by precipitation that is intense enough and of sufficient horizontal extent to fulfill the 37 GHz polarization criterion. Tables 2 and 4 show that not only are global morning totals 27% larger than evening totals, both for the January–February period and the July–August period, but the morning values are also higher for almost all latitude belts taken individually.

The same picture emerges when the data are considered by region: Table 3 shows that only in the Southern Midlatitudes are the evening totals larger than the morning totals, and then only by an insignificant amount. Elsewhere, the morning total is greater than the evening total by up to a factor of two. A significant morning excess during January–February is observed even in the SPCZ, where Albright et al. (1985) had inferred an afternoon maximum from their infrared satellite observations.

An obvious question to ask in the face of these consistent results is whether the apparent morning-evening differences could be an artifact of a systematic calibration fluctuation in the sensor itself. Since the number of pixels flagged would fall off rather rapidly as the 37 GHz polarization difference threshold is reduced below 15 K, even a modest calibration shift between morning and evening could conceivably give rise to the large observed differences in the totals.

Also, since we did not compute the number of pixels saved as a fraction of the total number of pixels falling within a particular geographic region, it is not yet clear whether any morning-evening biases might have been introduced by systematic differences in the sample size. However, we feel that the weight of the evidence speaks against either a calibration fluctuation or a sampling bias as being the source of the differences.

For example, the 37 GHz polarization thresholds which define various fractions of the archived pixels for selected regions were found to differ by at most 0.5 K from morning to evening, and usually much less. Differences of this size are consistent with random statistical fluctuations in the sample population. A calibration fluctuation of over 3 K would be required, however, to explain the mean observed difference of 44% between morning and evening totals for the tropics.

Perhaps even more significantly, 10th-percentile 85.5 GHz brightness temperatures (vertically polarized) are approximately 5 K lower

for the global AM pixels than for the PM pixels, suggesting that not only is there a larger fractional coverage by rain in the morning according to the 37 GHz polarization criterion, but the average intensity of convection within that rain is also greater. These results are the opposite of what would be expected if the true morning and evening characteristics of rainfall were in fact identical and a periodic calibration shift had simply biased the sample to include lighter rain events in the morning. Similarly, one would not expect such a large difference in the 85 GHz T_B if the total area sampled by the SSM/I were simply smaller during the evening passes due to missing data.

4. CONCLUSIONS

The data presented here for July–August 1987 and January–February 1988 appear to confirm a tendency toward significantly greater areal extent (based on the number of pixels meeting the 37 GHz polarization criterion) and greater mean intensity (based on the 85 GHz brightness temperature depressions of those pixels) of oceanic rainfall in the early morning (~0600 LST) than in the evening (~1800 LST), over much of the world's oceans. Only poleward of 45° latitude did we in some cases fail to observe a morning maximum. Interestingly, the morning totals were greater even in the South Pacific Convergence Zone, where Albright et al. (1985) had previously found an afternoon maximum, based on an IR brightness temperature threshold technique.

Because of the relatively crude method used here, the data presented must be regarded as preliminary. We plan to refine our approach in the near future, using improved algorithms and a larger SSM/I data set in order to reduce uncertainties related to sampling and in an effort to extract quantitative information concerning the total areal extent, physical characteristics, and temporal and regional variability of global oceanic precipitation.

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