Estimation of the Diurnal Cycle of Oceanic Precipitation from SSM/I Data

AWDHESH K. SHARMA

ST Systems Corporation, Lanham, Maryland

ALFRED T. C. CHANG

NASA/Goddard Space Flight Center, Greenbelt, Maryland

THOMAS T. WILHEIT

Department of Meteorology, Texas A&M University, College Station, Texas 77843

(Manuscript received 1 June 1990, in final form 11 March 1991)

ABSTRACT

A study of differences between the morning and evening monthly rainfall for 5° × 5° cells over the oceans from the SSM/I data has been conducted. The monthly rainfalls are estimated from the technique given by Wilheit et al. The difference between the morning and evening monthly rainfall arises due to the various random errors involved in the retrieval process, the sampling error in the observations, and the diurnal component of oceanic rainfall. The diurnal component is weak but clearly visible when averaged over large areas and for long time periods. The analysis shows that morning rainfall is consistently greater than evening rainfall. The Northern Hemisphere seems to have a larger diurnal variation than does the Southern Hemisphere. The maximum ratio between the morning and evening monthly rainfall is 1.7 while 1.2 is the more typical value.

1. Introduction

The diurnal variation of oceanic rainfall has been studied by several investigators in the past; For example, Meisner and Arkin (1987), Short and Wallace (1980), Riehl and Miller (1978), and Gray and Jacobson (1977) used satellite infrared data; Kraus (1963) used ship data and Kidder and Vonder Haar (1977) microwave data. Their studies are summarized in the next section. More recently a study by Shin et al. (1990a) showed that monthly area-averaged rain rates from low orbiting satellites will include a bias due to the existence of the semidiurnal cycle in the regions of SPCZ and ITCZ ranging from 5% to 10%. None of these studies has dealt with enough rainfall data to describe the phenomenon at a large scale in both space and time. Monthly rainfall estimates over the oceans from the Special Sensor Microwave Imager (SSM/I) measurements provide a basis for investigating the diurnal variation of oceanic rainfall at large scales both in space (globally between 50°N and 50°S) and time (1 yr).

The monthly oceanic rainfall values used here are estimated via a physically based technique that assumes a lognormal form for the rainfall intensity probability distribution function (PDF) (Wilheit et al. 1991). The PDF technique uses monthly histograms of vertically polarized brightness temperature on 5° × 5° cells at 19.35 GHz, 22.235 GHz, and a linear combination of the two.

The SSM/I was launched in June 1987 aboard the Defense Meteorological Satellite Program (DMSP) satellite, which was placed into a sun-synchronous near-polar orbit at an altitude of 833 km with an inclination of 98.8° and an orbital period of 102 min. Its equator crossing local times are at nearly 0600 and 1800 corresponding to ascending (northbound) and descending (southbound) portions of an orbit, respectively. The SSM/I is a microwave radiometer operating at 19.35, 22.235, 37, and 85.5 GHz and at both vertical and horizontal polarizations except 22.235 GHz where only vertically polarized radiation is measured. The spatial resolution of the SSM/I varies from 43 × 69 km^2 at 19.35 GHz to 13 × 15 km^2 at 85.5 GHz. The instrument measures over a swath of 1400 km. A detailed description of the SSM/I instrument is given by Hollinger et al. (1987). The data of the ascending and descending orbital segments were separately analyzed for investigating the oceanic rainfall variation with time.

The objective of this paper is to provide an accurate characterization that will contribute to an understanding of the physical processes associated with diurnal

Corresponding author address: Dr. A. K. Sharma, Code 975, Building 22, Room 301-b, NASA/Goddard Space Flight Center, Greenbelt Road, Greenbelt, MD 20771.

variability of oceanic rainfall. The results obtained show a strong evidence in favor of the existence of a modest diurnal cycle of oceanic rainfall. The nature of this diurnal cycle varies from region to region and from season to season. The magnitude and phase of this variability is important for estimating the bias involved in rainfall estimates that will be obtained from any sun-synchronous orbital platform. Another reason for this study is to provide a basis for comparison of general circulation model predictions of diurnal cycles with actual observations.

2. Past studies of rainfall diurnal cycle

(i) Three-hourly images of infrared radiation (IR) data from geostationary satellites were analyzed for a tropical convective precipitation study by Meisner and Arkin (1987). Their analysis of tropical oceanic data substantiates the diurnal cycle only in the convergence zones where near-noontime maxima were generally observed. They also noted substantial intra-annual fluctuations in the variances contributed by the diurnal cycle. They reported that the diurnal cycle over the tropical continents and the other continents during summer was much larger than that over the oceans. In virtually all areas, diurnal variance was largely explained by the first harmonic.

(ii) A study of outgoing IR values measured by scanning radiometers aboard the polar orbiting NOAA satellites in the water-vapor window (10.5–12.5 μm) was made by Short and Wallace (1980). Their analysis over the tropical oceans showed a three-tier structure with a pronounced morning bias in the frequency of high clouds and low clouds and a weak but statistically significant evening bias in the middle clouds. Their analysis over land and coastal waters supports the widely held view that cloudiness associated with convective activity is strongly modulated by boundary layer circulations that favor enhanced afternoon and evening cloudiness over islands and regions of elevated terrain and enhanced late night and morning cloudiness over coastal waters.

(iii) Differences between morning and evening temperatures of cloud tops over tropical continents and oceans were examined by Riehl and Miller (1978) using longwave radiation measurements made in the water-vapor window by the sun-synchronous satellite *NOAA-2* for the period 15 June–30 September 1974. They only considered the range of outgoing IR radiation that corresponds to high cloud tops such as might be associated with convective precipitation. They found that over land, Venezuela, and West Africa, the frequency of these very low values in outgoing IR was greater during the evening than during the morning, while over oceans, the Arabian sea and the GATE area, the reverse was true. They also reported that the absolute frequency of the very cold clouds was less over oceans than over land.

(iv) The existence of a widespread morning maximum in convective rainfall over the tropical oceans was also reported by Gray and Jacobson (1977) from a large number of stations on small, isolated tropical islands. They reviewed a number of papers on diurnal precipitation and reported that at many places heavy rainfall is 2–3 times greater in the morning than in the late afternoon–evening. They, however, noticed that the GATE observations showed a similar diurnal range in heavy rainfall, but the time of maximum occurrence was in the afternoon, about 6–7 h later than in most other oceanic regions. In this sense the GATE region may be considered as anomalous, which is likely a result of the particularly strong lower tropospheric vertical wind shear in this region and the frequency of squall and other line convection. They proposed that a morning maximum in the diurnal cycle in deep cumulus convection may result from day-versus-night variations in tropospheric radiational cooling between the weather system and its surrounding cloud-free region. They concluded that the more intense the deep convection and the more associated it is with organized weather systems, the more evident is the cycle.

(v) The study of maritime precipitation from convective and nonconvective clouds was made by Kraus (1963) from nine weather ships, in the Atlantic and Pacific oceans and 3-h July reports for the period 1950–61. He found that the frequency of nighttime (2100–0600 LST) shower precipitation is twice that in the daytime (0900–1800 LST) values. The results show that precipitation falls most frequently in the second half of the night. In fact, the midlatitude regions where rain is mostly nonconvective have the largest diurnal amplitude. The smallest amplitude is found near Bermuda, with predominantly convective rain. His analysis shows that the diurnal variation is related to the absorption of solar radiation. During the daytime, solar radiation absorbed by clouds should cause a reduction of the condensation rate in rising air and, therefore, a reduction of liquid-water content per unit volume. As a result, evaporation takes place at the tops of the clouds and the thickness of the clouds would tend to reduce. This acts to reduce precipitation. At night the tops of the clouds cool more than the cloud bases. The resulting instability produces more vertical overturning. This results in the raising of the quasi-permanent inversion height that allows deeper convection and more rainfall.

(vi) A study of oceanic precipitation, from the microwave brightness temperatures obtained from *Nimbus-5* for the period December 1972–February 1973 was made by Kidder and Vonder Haar (1977). In their study, the zonal mean freezing levels derived by Oort and Rasmusson (1971) and Taljaard et al. (1969) were used. Their results show that the difference between the local noon and the local midnight frequency is small, but the heavier rainfall rates tend to occur more frequently near local noon. Also the ratios of the frequencies of light, moderate, and heavy rain were seen to be relatively constant over the tropical oceans. The small difference they found between noon and midnight is compatible with the present results of maximum and minimum being near morning and evening, respectively.

3. Methodology

The present analysis is principally based on the monthly oceanic rainfall inferred from the monthly histograms of a linear combination of brightness temperatures, (2TB19 − TB22). This combination of brightness temperatures was chosen in order to reduce the water-vapor effect in the retrievals of rainfall that can introduce errors in the determination of the rain intensity at low rain rates (Wilheit 1986). The oceanic area between latitudes of 50°N and 50°S was initially divided into 5° × 5° latitude–longitude grid boxes. At each grid box the SSM/I brightness temperatures were collected to form a histogram with 1-K bins ranging from 100 to 295 K over a 1-month period. A typical histogram shows a skewed distribution that has a long tail at higher brightness temperatures (Kedem et al. 1990), suggesting that it is composed of two distributions. The first part is a roughly normal distribution of colder brightness temperatures of nonraining atmospheres over the highly reflective oceanic background, and the second part is due to raining atmospheres contributing the warmer brightness temperatures.

An algorithm has been developed by Wilheit et al. (1991) to estimate the monthly total precipitation from the brightness temperature histograms. The procedure assumes a lognormal form for the rainfall PDF and assumes that the nonraining portion of the histogram can be approximated by a normal distribution of brightness temperatures. The algorithm solves iteratively for three parameters of the lognormal distribution and for the mean and variance of the nonraining portion of the histogram. The rainfall totals are then calculated from the lognormal parameters and the time period covered by the histogram. The algorithm assumes a relationship between the brightness temperature and the rain rate (a R–T relationship) based on Wilheit et al. (1977). An important parameter for the R–T relationship is the freezing level. This algorithm estimates the freezing level by fitting the 19V and 22V scattergrams. A similar approach of rainfall retrievals has been used by Shin et al. (1990b) using the single frequency (19.35 GHz) *Nimbus-5* ESMR data. The freezing level was estimated from the climatological data in their study.

The R–T relationship is highly nonlinear, and thus, direct application of the transfer function will result in an underestimate of the rainfall rate, if there is any inhomogeneity in the field of view (FOV). This error due to inhomogeneity of the rain field in a FOV of microwave radiometer is known as the beam-filling problem and has been extensively studied by Short and North (1990), and Chiu et al. (1990). They concluded that a beam-filling correction factor k can be found ranging from 1.5 to 2.5 in the GATE (GARP Atlantic Tropical Experiment) region. It is not possible to correct globally for this bias on the basis of this limited study in space and time. As a matter of fact, the variation of k with respect to space and time remains an open question. Therefore, no correction has been made for this bias in the present study. It is unlikely that there is a significant diurnal variation in the beam-filling error. Therefore, as long as we are concerned with percentage change in rainfall, rather than absolute amounts, the beam-filling error will not be a factor.

The Wilheit et al. (1991) algorithm estimates the freezing level by using the 19V and 22V histograms. Since this approach could be very sensitive to instrument calibration errors, which, in turn, could have a significant diurnal cycle because of the solar influence on the spacecraft and instrument, we have chosen to use an average of the freezing levels computed from the morning and evening histograms for each location. The remainder of the algorithm is much less sensitive to calibration errors.

4. Data analysis and results

The variation of local solar time (LST) with latitude is shown in Fig. 1 for both the ascending and descending segments of the orbit. The range of LST varies from 0610 to 0735 at 50°S; 0540–0635 at the equator; and 0445–0610 at 50°N for the ascending segment of the orbit; and from 1810 to 1935 at 50°N; 1740–1835 at the equator; and 1645–1810 at 50°S for the descending segment of the orbit. The maximum difference of LST for each of these segments between 50°N and 50°S is 2 h 50 min. The difference between successive ascending and descending LSTs at any given point on the earth is at least 9 h 10 min. The observations are slightly biased towards the nighttime at 50°N and towards the daytime at 50°S. We shall refer to rainfall data obtained from the ascending portion of the orbit as morning rainfall and rainfall data obtained from the descending portion as evening rainfall. After such categorization our goal here is to investigate the difference between the morning and evening rainfall on a monthly basis.

Ten months of data from the period August 1987–November 1987 and February 1988–July 1988 have been analyzed. The instrument was turned off for a large fraction of the two missing months (December 1987 and January 1988). Although the present estimates of monthly rainfall might be inaccurate due to biases in the retrieval process such as the lack of a beam-filling correction, the relative differences will not suffer this error. The resulting morning-minus-evening monthly rainfall differences were mapped for each of these ten months of data.

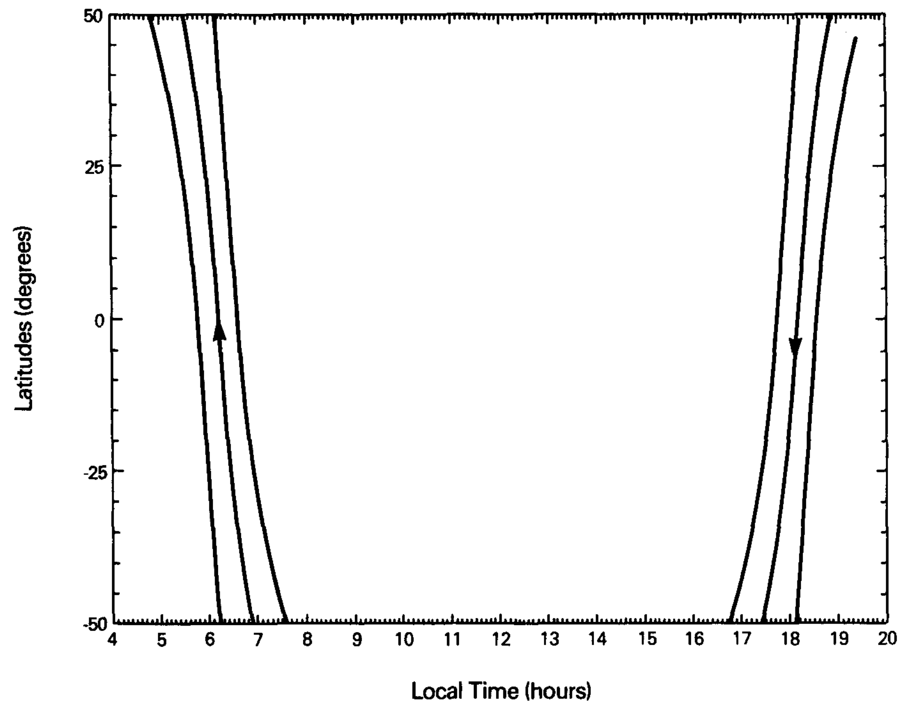

FIG. 1. Variation of local solar time (LST) with latitude for ascending and descending segments of an orbit.

On the basis of these maps, it was possible to identify the broad geographical regions where the variation was small. However, it was not possible to categorize all the geographical regions in which the systematic patterns of the negative and positive values of the difference can be seen because the random uncertainty in the rainfall totals is comparable to or larger than the diurnal difference. When the total number of pixels of the difference of opposite signs were compared, it was found that the number of pixels with a positive difference (morning − evening) of rainfall is greater than that of the negative difference. This suggests that the total area of positive difference is more than that of the total area of negative difference and that the morning rainfall is more than the evening rainfall over the oceanic area of the study.

The differences of morning minus evening monthly rainfall at each grid box were examined on a histogram for determining the character of the frequency distributions of each of these ten months of data. A sample histogram, Fig. 2, shows the percentage frequency distribution for the total ten months of data together where the abscissa shows the upper limit of the difference of (morning − evening) monthly rainfall (mm month^{-1}) and the ordinate shows the percentage frequency. All histograms appear to be roughly normally distributed with some skewness toward left of the origin and a maximum at 5 mm month^{-1} with two peaks at 200 and −200 mm month^{-1} at extreme ends. The extreme interval widths are about ten times larger than the interval size of the histogram, however, the extreme left peak is three times larger than the extreme right, which indicates that the difference of morning-minus-evening monthly rainfall is much larger than the difference of evening minus morning monthly rainfall. It is also ev-

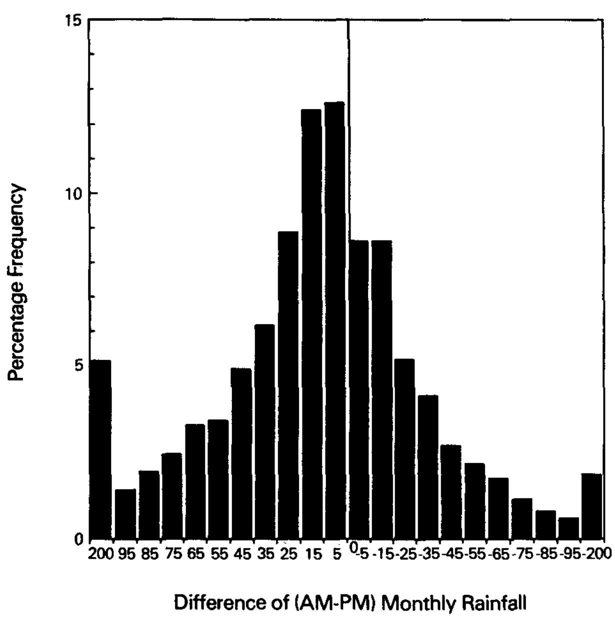

FIG. 2. Percentage frequency distribution of the morning-minus-evening monthly rainfall for a total ten months of data.

ident from this figure that the area to the left side of the origin is about 60% of the total area.

For all ten months of data, total monthly rainfall values for morning and evening times at each grid box were ordered, and grouped at each ten percentile for each month. Figure 3 shows as an example of such a comparison for a total of ten months of data where the ordinate shows the logarithmic monthly rainfall and the abscissa shows the percentile of data. It is evident from this figure that the morning rainfall seems to have consistently greater value than the evening rainfall. It should be noticed that for up to 90% of the data, the values of morning and evening monthly rainfall are less than 200 mm month^{-1} and their differences are less than 25 mm month^{-1}. The highest ten percentile of data may include the least reliable estimates as a result of saturation occurring at high rainfall intensity or possible erroneous data and, therefore, should be taken somewhat less seriously. The maximum morning and evening monthly rainfall values were 731.4 and 596.3 mm month^{-1} observed, and their difference 135.1 mm month^{-1} was found for the total ten months of data. One should be very careful about this representation because it is very sensitive to the maximum values obtained for the two times, even a single bad maximum value in either time can show the difference, not an actual characteristic of the percentile.

Figure 4 depicts the percentile distribution of morning-minus-evening normalized percentage cumulative monthly rainfall for each of these ten months of data. Mathematically it can be written as

$$100\left(\frac{\sum_{i=1}^{n_j} X_i}{\sum_{i=1}^{N} X_i} - \frac{\sum_{i=i}^{n_j} Y_i}{\sum_{i=1}^{N} Y_i}\right)$$

where n_j is the number of points for jth percentile, N is the total number of points, and X_i and Y_i are the morning and evening monthly rainfall at each grid box arranged in an increasing order. The maximum positive difference (morning − evening) 4.36 is found at the 80th percentile for March 1988, while the maximum negative difference 1.68 is found at 80th percentile for November 1987. The negative difference is mainly observed for the month of November 1987, which may be interpreted as few data points in the Southern Hemisphere associated with convective activity during the evening time of the descending pass. However, September 1987 also shows small negative differences in the higher percentiles caused by the evening, higher values of monthly rainfall as observed in the intertropical convergence zone (ITCZ) around the equator. These cases of a negative difference underscore the sensitivity of the observations to sampling error.

In the following sections, we will discuss the seasonal and regional variations of morning minus evening monthly oceanic rainfalls. These morning–evening differences were properly normalized and then meridional and zonal averages were computed. The average of data lying between the latitudes of 50°N and 50°S is called the meridional average. The data were further reduced by averaging over 20° bands of latitude and longitude. From our data analysis, it was noticed that almost all months show a positive difference except for the midlatitude band, −30° to −50° in the Southern Hemisphere, where the annual change (from August 1987 to July 1988) in the difference decreases and the negative differences were observed only in August and November 1987. An increase in annual change in the diurnal cycle from the Southern to the Northern Hemisphere is also observed. In the Northern Hemisphere August 1987 shows the maximum diurnal difference of about 50% in the midlatitude band, +50° to +30°, and a second maximum is shown by July 1988. In the tropics and the midlatitude regions of the Northern Hemisphere these months generally had the highest rainfall. February 1988 shows the minimum difference as well as minimum monthly rainfall in all bands. The maximum diurnal variation and its annual variation occurs between 280° and 300°E, which has a small monthly rainfall and, therefore, can be attributed to sampling error. Similarly the maximum negative difference occurs between 20° and 40°E, which has a minimum number of grid points and, therefore, should be taken less seriously. The minimum 20% an-

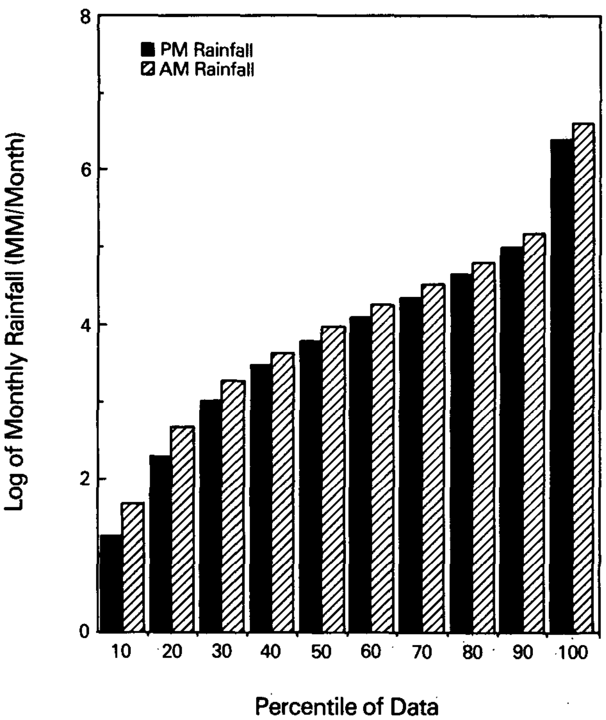

FIG. 3. Logarithmic distribution of morning and evening monthly rainfall (mm month^{-1}) for a 5° × 5° cell over the oceans for a total of ten months of data.

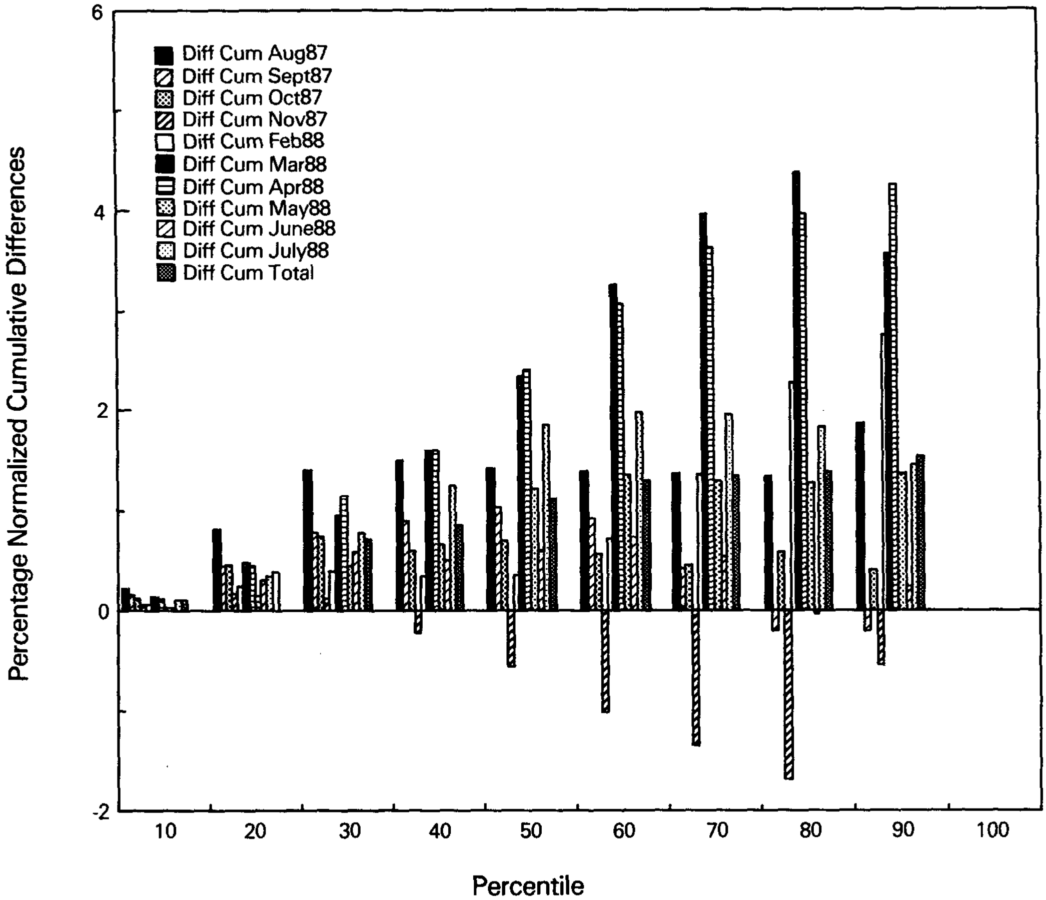

FIG. 4. Percentage differences of morning and evening normalized cumulative monthly oceanic rainfall for each ten months of data.

nual change in diurnal cycle is observed in the 0°–20°E longitude band. A large annual variation in the diurnal cycle can be attributed to sampling fluctuations caused by our short (1-month) averaging period and small averaging area (20° bands of latitude and longitude).

To reduce the influence of sampling fluctuations, averages over three months were computed. These months were based on four seasons of a year as shown in Figs. 5 and 6, respectively, where (S + O + N) means September, October, and November 1987, (F + M + A) means February, March, and April 1988; (M + J + J) means May and June 1987 and July 1988. In these figures almost all the data lie in the positive region of the (morning − evening) difference, which implies that there is a pronounced morning bias in the seasonal oceanic rainfall. This bias is weak but important (ranging from 1.2 to 1.7) and should not be ignored in any scheme estimating the average of rainfall over time and space from satellite data not having a uniform temporal coverage. It can be seen in the Fig. 5 that the order of the months from minimum to maximum diurnal change is approximately reversed when moving from the Northern Hemisphere to the Southern Hemisphere, as expected. The maximum rainfall occurs during their respective summer months. For the data (S + O + N), (D + J + F), and (M + A + M), the diurnal cycle increases from 50°N to the equator and then decreases from the equator to 50°S with a maximum change attained by the (M+A+M) at the equator. The (J + J + A) data shows a sinusoidal variation from 50°N to 50°S with a minimum attained by (S + O + N) at the equator. Meridional averages of the differences as shown in Fig. 6 are consistent with the rainfall patterns generally observed in the corresponding 20° longitudinal bands, and a large scatter in some of these bands is associated with the lesser number of points in those regions.

5. Conclusions

We have compared morning and evening monthly oceanic rainfall retrieved from the SSM/I data using an algorithm developed by Wilheit et al. (1991). This algorithm produces rainfall indices rather than the estimates of rainfall rates because of uncertainty in the calibration. The model used for calculating the monthly rainfall indices is based on many assumptions and approximations that are physically reasonable. However, a beam-filling correction has not been applied to the monthly rainfall and, therefore, it is called an index rather than a measurement. It is unlikely that there is a significant diurnal variation in the beam-filling correction factor. Since we are concerned with percentage change in rainfall, rather than absolute amounts, the beam-filling correction is not a factor. The difference between the morning and evening monthly rainfall arises due to both the various random errors involved in the retrieval process and the diurnal component of oceanic rainfall. The random errors are reduced when averaged over a large area (20° bands of latitude and longitude meridional and zonal averages) and over a long period of time (three months). Although the diurnal component is weak, it can clearly be seen when the data are averaged over a large area and for a long time period. The histograms of brightness temperatures from which these values of monthly rainfall were derived also show longer tail toward higher brightness temperatures for morning than for the evening. In order to minimize the number of sources for random errors, the same freezing level has been assumed for both the morning and evening cases at each location. The local solar time difference between the morning and evening observations is at least 9 h, which is shorter toward nighttime in the Northern Hemisphere and toward day time in the Southern Hemisphere. The Northern Hemisphere seems to have larger diurnal variation than the Southern Hemisphere. This can be explained on

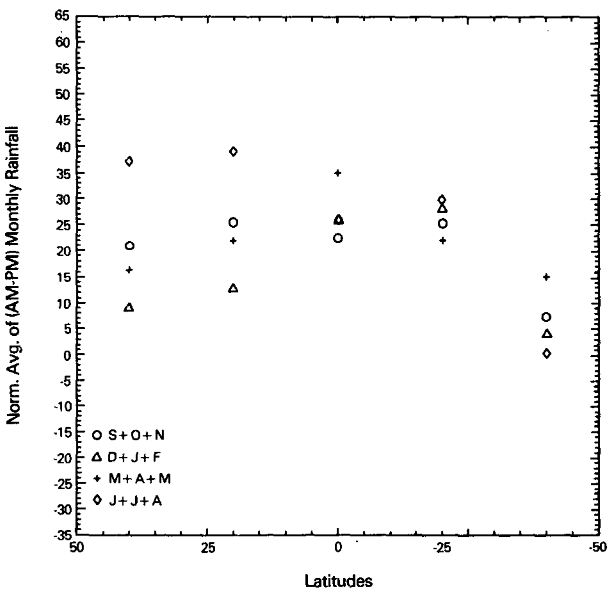

FIG. 5. The variation of zonal and seasonal (3-month averages) normalized differences of morning and evening monthly rainfalls with 20° bands of latitudes.

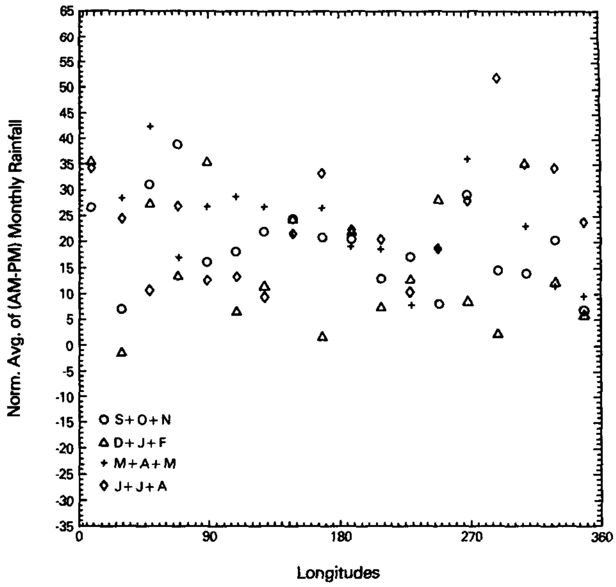

FIG. 6. The variation of meridional and seasonal (3-month averages) normalized differences of morning and evening monthly rainfalls with 20° bands of longitudes.

the basis of the diurnal phase. If the rainfall maxima is shifted back from 0600 LST then the Northern Hemisphere measurements would see it better than the Southern Hemisphere measurements because of their timing. Another reason for the difference of the diurnal cycle in these hemispheres could be due to different proportions of land and sea but we do not know the mechanism for that. The maximum diurnal variation as well as maximum annual change in diurnal variation is observed between 280° and 300°E, which has a minimum number of samples and, therefore, may not be statistically significant. Also, the maximum negative change in the diurnal cycle occurs between 20° and 40°E, which has very few observations and may be attributed to sampling errors. When the sample size is increased by averaging three months of data, then a consistently pronounced morning bias in the monthly oceanic rainfall is found. This bias is not nearly as large as the 2:1 or 3:1 suggested by other authors but the maximum 1.7:1 with 1.2:1 is more typical.

Acknowledgments. The work presented here was carried out at the Goddard Space Flight Center, NASA, Greenbelt, Maryland, while the first author was a NASA-NRC resident research associate.

REFERENCES

Chiu, L. S., G. R. North and D. A. Short, 1990: Rain estimation from satellites: Effect of finite field of view. *J. Geophys. Res.*, **95,** 2177–2185.

Gray, W. M., and R. W. Jacobson Jr., 1977: Diurnal variation of deep cumulus convection. *Mon. Wea. Rev.*, **105,** 1171–1188.

Hollinger, J., R. Lo, G. Poe, R. Savage and J. Peirce, 1987: *Special Sensor Microwave/Imager User's Guide*. Naval Research Laboratory Washington, D.C., 120 pp.

Kedem, B., L. S. Chiu and G. North, 1990: Estimation of mean rain rates: applications to satellite observations. *J. Geophys. Res.*, **95**, 1965–1972.

Kidder, S. Q., and T. H. Vonder Haar, 1977: Seasonal oceanic precipitation frequencies from Nimbus 5 microwave data. *J. Geophys. Res.*, **82**, 2083–2086.

Kraus, E. B., 1963: The diurnal precipitation change over the sea. *J. Atmos. Sci.*, **20**, 551–556.

Meisner, B. N., and P. A. Arkin, 1987: Spatial and annual variations in the diurnal cycle of large-scale tropical convective cloudiness and precipitation. *Mon. Wea. Rev.*, **115**, 2009–2032.

Oort, A. H., and E. M. Rasmusson, 1971: Atmospheric circulation statistics. NOAA Prof. Pap. 5, United States Department of Commerce, Rockville, Md, 323 pp.

Riehl, H., and A. H. Miller, 1978: Differences between morning and evening temperatures of cloud tops over tropical continents and oceans. *Quart. J. Roy. Meteor. Soc.*, **104**, 757–764.

Shin, K. S., G. R. North, Y. S. Ahn and P. A. Arkin, 1990a: Time scales and varibility of area-averaged tropical oceanic rainfall. *Mon. Wea. Rev.*, **118**, 1507–1516.

———, P. E. Riba and G. R. North, 1990b: Estimation of area-averaged rainfall over tropical oceans from microwave radiometry: A single channel approach. *J. Appl. Meteorol.*, **29**, 1031–1042.

Short, D. A., and J. M. Wallace, 1980: Satellite-inferred morning-to-evening cloudiness changes. *Mon. Wea. Rev.*, **108**, 1160–1169.

———, and G. R. North, 1990: The beam filling error in the Nimbus 5 Electronically Scanning Microwave Radiometer observations of global atlantic tropical experiment rainfall. *J. Geophys. Res.*, **95**, 2187–2193.

Taljaard, J. J., H. van Loon, H. L. Crutcher and R. L. Jenne, 1969: *Climate of the Upper Atmosphere, Part 1, Southern Hemisphere.* vol. 1. National Center for Atmospheric Research, National Weather Records Center, and Department of Defense, 134 pp.

Wilheit, T. T., A. T. C. Chang and L. S. Chiu, 1991: Retrieval of monthly rainfall indices from microwave radiometric measurements using probability distribution functions. *J. Atmos. Oceanic Technol.*, **8**, 118–136.

——, 1986: Some comments on passive microwave measurement of rain. *Bull. Amer. Meteor. Soc.*, **67**, 1226–1232.

——, A. T. C. Chang, M. S. V. Rao, E. B. Rodgers and J. S. Theon, 1977: A satellite technique for quantitatively mapping rainfall rates over the oceans. *J. Appl. Meteor.*, **16**, 551–560.