SATELLITE METEOROLOGY

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Contents

Summary

This article can largely be divided into two parts: 1, a brief review on the theoretical bases for satellite remote sensing of various meteorological parameters, and 2, several selected applications of satellite observations to tropical meteorology. In the theoretical review, the solutions to the radiative transfer equations are described under a variety of conditions, in relation to the physical principles by which certain meteorological parameters can be retrieved. That is, under clear-sky conditions, surface temperature and atmospheric temperature and moisture profiles may be retrieved, respectively, from window and sounding channels in infrared and/or microwave spectrum. Cloud properties, such optical depth and effective radius, can be retrieved from reflected solar radiation at visible and near-infrared channels. Observations in the microwave spectrum can be used to infer such meteorological variables as precipitable water, cloud liquid water, cloud ice water, rainfall, and ocean surface wind speed. Additionally, active sensors, such as cloud and precipitation radars, can provide the vertical distribution of hydrometeors. Scatterometers, on the other hand, can measure both ocean surface speed and direction. Three topics are selected to demonstrate the applications of satellite observations in the tropics: hurricane studies, tropical rainfall and energy exchange at ocean surface. Studies on tropical cyclone's intensity, structure and water balance are first introduced. For tropical rainfall, works primarily from Tropical Rainfall Measuring Mission satellite observations are reviewed, including rainfall climatology, horizontal and vertical structures of rain fields, diurnal variation of rainfall and latent heating related to precipitation. Finally, topics on deriving radiative and turbulent fluxes at ocean surface are discussed. Since clouds are an important factor in influencing these fluxes, we also introduced research on satellite-derived cloud properties, with an emphasis on the efforts made by the International Satellite Cloud Climatology Project.

1. Introduction

What happens in the tropical region is vital for understanding and forecasting the global weather and climate since the region provides the "fuel" to the engine of the atmospheric general circulation. However, the vast majority of the tropics are covered by oceans where surface-based observations of the atmosphere and surface are sparse. This makes satellite-based remote sensing an indispensible part of the tropical observation system. Since the first satellite that was completely dedicated to satellite meteorology - TIROS 1 (Television and Infrared Observational Satellite) – was launched in 1960, satellite observation has been playing an increasingly important role in the field of tropical meteorology. An historical overview of important milestones in satellite meteorology is given by Kidder and Vonder Haar (1995). While interested readers may consult that book for details, there are still several developments worth mentioning here: First, since the mid of 1960s, meteorological satellites have become operational (as opposed to experimental) in the United States; their coverage became continuous in time and almost anywhere on globe in space. One important implication of this routine observation is that no more tropical cyclones will go undetected any more, which is a plausible explanation that the number of recorded Atlantic storms jumped in the 1960s. Second, the availability of microwave sensors, particularly SSM/I (Special Sensor Microwave/Imager) since late 1980s, dramatically improved the quantitative estimation of global precipitation. Before this microwave era, rainfall estimates in the tropical oceans were primarily relied on the few measurements from commercial ships and sparsely located atolls; there was no clear consensus on the climatology of tropical rainfall amount. Third, satellite meteorology leaped a large step forward with the introduction of active sensors. The first precipitation radar on TRMM (Tropical rainfall Measuring Mission) and the first cloud radar on CloudSat have proved to be extremely valuable in understanding the vertical structures of clouds and precipitation.

This article is in no means to be inclusive of all the important developments in satellite meteorology. Instead, the author decided to focus on two things: for laying the physical ground of remote sensing in the first half and then in the second half "cheery-picking" several applications of satellite observations, to which the author deemed important to tropical meteorology. For interested readers, the following books may be found helpful: Stephens (1994), Kidder and Vonder Haar (1995), and Liou (2002). Additionally, the online materials published by University Corporation for Atmospheric Research (UCAR) COMET (Cooperative Program for Operational Meteorology, Education and Training) program at the website: http://www.meted.ucar.edu/tropical/textbook/ch3/Ch3_Remote_Sensing.pdf are also very informational. Remote sensing is the inverse process of radiative transfer. To perform retrieval of meteorological parameters from satellite measurements, we first have to understand how radiative energy at certain wavelength propagates in the atmosphere under given conditions. Therefore, this article starts in Section 2 with the basic theoretical descriptions of radiative transfer equations, followed by its solution under a specific condition and how the solution is relevant to remote sensing of certain meteorological parameters. In Section 3, we focused on three topics of applications of satellite remote sensing: hurricanes, precipitation and air-sea energy exchange (including the influence by clouds). The reasons to pick up these three topics for tropical meteorology are explained at the beginning of each section. One topic – atmospheric sounding using satellite observations – is not given much space in this article, although it is a very important topic, especially for improving numerical weather prediction models. Radiances from sounding channels have been assimilated in many operational forecasting models since early 1990s. Interested reads may consult Eyre et al. (1993) for more details.

2. Principles of Satellite Remote Sensing

Meteorological satellites have been designed so far to fly either polar or geostationary orbit. A geostationary satellite at 36,000 km altitude over the equator has a traveling speed equaling the angular velocity of the Earth, so that its position relative to an observer on the Earth appears to be fixed. This gives the ability for a sensor on a geostationary satellite to measure a target on the Earth in highly frequent repetition. For example, we can measure a tropical cyclone by a radiometer on a geostationary satellite every 15 minutes to monitor its evolution. However, its high altitude prevents sensors that have large field of view (for example, current microwave radiometers) from being placed on a geostationary satellite in order to obtain a reasonable spatial resolution of measurements. A polar orbiter, on the other hand, overflies higher latitudes on the Earth, normally at an altitude below 2000 km. While the satellite's orbital plane is fixed in space, with the rotation of the Earth, sensors on the polar orbiting satellite can observe different parts of the Earth. A target on the Earth usually can be observed twice daily by a typical meteorological satellite.

Remote sensing refers to as measuring certain features of a target without being in direct contact. It can be divided into two categories – active and passive sensing – depending on whether or not the sensor-received energy is first transmitted from the sensor. A satellite-borne radiometer receives radiative energy from the Earth and its atmosphere. The received energy may be the emission originated from, or the reflection by the Earth and atmosphere, or a combination of both. Therefore, the radiometer-received energy contains the information on the characteristics of the atmosphere and its underlying surface. The task of satellite remote sensing is then to extract meteorologically useful information from the satellite-received radiation; which is termed as "retrieval problem", or, "backward problem". In other words, the information we have (satellite received energy) is a result of complicated interactions between electromagnetic waves and matters in the sensor's viewing volume. Yet, the goal of the retrieval problem is to infer one certain feature of these matters. For example, we are to retrieve rainfall from observations at several microwave channels. The radiation received by the satellite is a combination of surface emission and reflection, emission by atmospheric gases, emission and scattering by hydrometeors at all levels of the atmospheric column. Rain drops near the surface are only one of the many factors that contribute to the upwelling radiation reaching to the satellite. The same surface rainfall rate could result in very different radiative energy under different surface conditions or with different atmospheric/cloud structures. On the other hands, corresponding to the same satellite received radiation, surface rainfall could have many possible intensities depending again on the surface and atmospheric conditions. Therefore, retrieving meteorological parameters by satellite remote sensing is intrinsically an ill-posed problem. To obtain a quality result, a retrieval algorithm developer is generally required to (1) select sensor channels that are most sensitive to the parameter to be retrieved, and (2) use *a priori* knowledge as constraints to the retrieval algorithm so that only the most probable solution is to be inferred from the observations.

Figure 1 shows the atmospheric transmittance for wavelength from 0.1 μm to 3 cm. Transmittance is defined varying from 0 to 1. A transmittance close to 1 indicates that an electromagnetic wave at the wavelength can travel through the atmosphere with little absorption (window region). Otherwise, a transmittance close to 0 indicates that the atmosphere at the wavelength is nearly opaque. What wavelength to select for a retrieval algorithm depends on the parameters intended to infer. For example, to retrieve surface feathers (e.g., sea surface temperature, surface wind), it is preferred to avoid the atmospheric influence as much as possible; thus, window wavelength (e.g., 11-12 μm) is preferred. On the other hand, to retrieval atmospheric temperature profile, it is required that the satellite received radiances at several wavelengths originated primarily from different atmospheric levels; thus wavelengths near (O_2 or CO_2) absorption lines (e.g., 15 μm, 0.5 cm) should be selected. It is particularly noted that of the broad spectrum of wavelengths used in satellite radiometry, microwaves convey especially rich information because of their ability to penetrate deep cloud and precipitation layers, so that information beneath the cloud can be accessed from the satellite. For this reason, satellite microwave data have been increasingly utilized in retrieving cloud water and rainfall intensities and in improving numerical weather forecasts, particularly under raining conditions.

2.1. Radiative Transfer in the Atmosphere

While the goal of satellite meteorology is to determine meteorological feature from received radiation, i.e., the backward problem, it cannot be done without first understanding how much radiation would reach to the satellite given a known surface and atmospheric condition, i.e., the forward problem. The radiative energy received by a satellite radiometer is either emitted by the earth-atmosphere system (terrestrial radiation) or first originated from the sun and the reflected by the earth-atmosphere system (solar radiation). Approximately, 99% of terrestrial radiative energy is distributed in the spectra with wavelength longer than 4 μm, while 99% of solar energy is with wavelength shorter than 4 μm. Therefore, in radiative transfer computations, it is common to treat terrestrial and solar radiations separately. For terrestrial radiation, the thermal emission is the fundamental source of the energy while radiation from the sun is ignored (unless the sensor is directly pointing to the sun). For solar radiation, the radiation from the sun is the fundamental source of energy while emission from the earth-atmosphere system is ignored.

The radiative transfer equation for an unpolarized wave at a given wavelength under a plane-parallel atmosphere can be expressed by

$$\mu\frac{dI(\tau,\mu)}{d\tau} = I(\tau,\mu) - \frac{\omega_0}{2}\int_{-1}^{+1} I(\tau,\mu,\mu')P(\cos\Theta)d\mu' - (1-\omega_0)B(\tau)\,, \tag{1a}$$

for terrestrial radiation, and

$$\mu\frac{dI(\tau,\mu)}{d\tau}=I(\tau,\mu)-\frac{\omega_0}{2}\int_{4\pi}I(\tau,\mu,\mu')P(\cos\Theta)d\mu'-\frac{\omega_0}{4\pi}F_0e^{-\tau/\mu_0}P(\cos\Theta_0), \tag{1b}$$

for solar radiation. In the equations, τ is optical depth defined to be zero at the top of the atmosphere and increases with lowering altitude, μ is the cosine of the emergent zenith angle of the wave, $I(\tau, \mu)$ denotes the radiance at optical depth level τ and emergent direction μ, ω_0 is single-scatter albedo, Θ is the angle between radiation's emergent direction μ and incident direction μ', Θ_0 is the angle between radiation's emergent and sunlight direction, and F_0 is solar flux at the top of the atmosphere. $B(\tau)$ is the Planck's function at optical depth τ. Several methods have been developed for solving the radiative transfer equations, including discrete ordinate method and doubling-adding method, etc. Here we are not going to describe these methods although computed results using them will be shown in later sections. Interested readers may refer to radiation textbooks, e.g., Liou (2002), for detailed descriptions. Instead, we are going to simplify the equation under idealized conditions, in an effort to grasping the physical insight by only extracting the primary signature without regarding to the quantitative details.

2.2. Thermal Radiation under Clear-Sky

At infrared and microwave wavelengths under clear-skies, the scattering (by air molecules and aerosols) is negligible (ω_0=0). Eq.(1a) can be rearranged as

$$\mu\frac{dI(\tau,\mu)}{d\tau}=I(\tau,\mu)-B(\tau) \quad . \tag{2}$$

The solution for upwelling radiation at the top of the atmosphere is then

$$I(0,\mu)=I(\tau^*,\mu)\exp(-\frac{\tau^*}{\mu})+\int_0^{\tau^*}B(t)\exp(-\frac{t}{\mu})\frac{dt}{\mu}, \tag{3}$$

where τ^* is the atmospheric total optical depth measured at ground. Defining a transmission function $\Im(t,0,\mu)=exp(-t/\mu)$ and a weighting function $W(t,0,\mu)=d\Im(t,0,\mu)/dt=-\frac{1}{\mu}\exp(-t/\mu)$, (3) becomes

$$I(0,\mu)=I(\tau^*,\mu)\Im(\tau^*,0,\mu)+\int_{\tau^*}^{0}B(t)W(t,0,\mu)dt\,. \tag{4}$$

The solution states that the radiance received by a satellite sensor consists of the contributions of transmitted portion of surface emission and the weighted sum of the atmospheric emission at different levels. This equation becomes important as we retrieve surface temperature and atmospheric temperature and moisture profiles. At window wavelengths the atmospheric contribution is minimal, the main contribution will be from the surface (the first term on right side of the equation), which allows us to derive surface temperature at infrared window channels under clear-skies. On the other hands, if we choose absorbing channels that have weighting functions peaked at different levels, we may combine the observations at these channels to derive atmospheric profiles. A special circumstance is thermal infrared window channel observation over overcast sky, in which cloud top acts the same as a surface, so that the satellite observed signal indicates the cloud top temperature.

2.2.1 Surface Temperature Retrieval

The principle of retrieving surface temperature from measurements of brightness temperatures at split-window (11 and 12 μm) channels is based on Eq.(4). At atmospheric window channels, the absorption by atmospheric gases (primarily water vapor) is minimal, we may further simplify Eq.(4), and express the satellite received radiance at channel *i* as

$$I_i = B_i(T_s)\Im_i + B_i(T_a)(1-\Im_i), \tag{5}$$

where T_s and T_a are surface and atmospheric mean temperatures, respectively. The transmission function $\Im_i = \exp(-\int k_i du) \approx 1 - k_i u$, where k_i is the water vapor absorption coefficient and u is the water vapor path. If we use i=1 and 2 to denote the split-window channels of 11 and 12 μm, respectively, surface temperature may be approximately expressed by [interested readers may refer Liou (2002) for detailed derivation]:

$$T_s = T_{B1} + \eta(T_{B1} - T_{B2}), \tag{6}$$

where $\eta = \dfrac{k_1}{k_2 - k_1}$, and T_{B1} and T_{B2} are the brightness temperatures observed at 11 and 12 μm channels, respectively. Based on this principle, empirical methods by regressing brightness temperatures observed by NOAA satellite AVHRR (Advanced Very High Resolution Radiometer) split-window channels to buoy-measured sea surface temperatures (SSTs) have been developed (e.g., McClain et al., 1985).

2.2.2 Atmospheric Sounding

In the thermal infrared and microwave spectra, atmospheric molecules absorb radiation by rotational or vibrational-rotational transition, which occurs at certain absorption lines (or bands) with strength and width specific to the gas species. For example, CO_2 has a strong absorption line centered 15 μm and water vapor has a strong line centered 6.3 μm wavelength. Around the absorption line, absorption does not occur at a single wavelength, but rather spreads a wavelength range following a line shape function. Mass absorption coefficient at frequency ν, k_ν, may be approximately described by a Lorentz model (Stephens, 1994):

$$k_\nu = \frac{S\tilde{p}\alpha_{L0}/\pi}{(\nu-\nu_0)^2 + \alpha_{L0}^2\tilde{p}^2}, \tag{7}$$

where ν is the frequency in consideration and ν_0 is the frequency at the absorption line center. α_{L0} is the Lorentz half-width, frequency shift corresponding to half power point of the line, at pressure p_0=1 atmosphere. S is the line strength, and $\tilde{p} = p/p_0$ is the pressure normalized by p_0. Eq.(7) indicates that the line shape widens with atmospheric pressure, called pressure broadening, so that the absorption is more concentrated near the line center in the upper atmosphere while it expands more toward the wings in the low atmosphere. Consider that the upwelling radiation is measured at multiple wavelengths from the line center to its wings for the absorption line of a vertically well-mixed gas (e.g., CO_2, O_2, …). The radiation measured near the line center corresponds to the temperature of upper atmosphere while radiation measured at the far wing corresponds to the temperature of the lower atmosphere; therefore, the atmospheric temperature profiles may be derived using an inversion algorithm.

Let us derive an approximate expression of the weighting function as a function of height for a vertically well-mixed gas that has a mixing ratio of w_g. The optical depth of a layer between pressure level p_1 and p_2 ($p_2>p_1$) in an atmosphere with hydrostatic balance may be written

$$\tau = \frac{w_g}{2p_0 g}\int_{p_1}^{p_2}\frac{Sp\alpha_{L0}/\pi}{(\nu-\nu_0)^2+\alpha_{L0}^2\tilde{p}^2}dp\,. \tag{8}$$

To simplify the equation, let us consider the absorption in the line wings where $\nu-\nu_0>\alpha_{L0}\tilde{p}$, so that optical depth at pressure level p equals

$$\tau \approx \frac{S\alpha_{L0}w_g}{gp_0(\nu-\nu_0)^2}p^2\,. \tag{9}$$

Assuming that atmospheric pressure decreases with height following a scale height H, i.e., $p(z)=p_0e^{-z/H}$, then optical depth at altitude level z,

$$\tau(z)=\tau^* e^{-2z/H}\,, \tag{10}$$

and τ^* is the optical depth for the entire atmosphere. From the definition of weighting function,

$$W(z)=\frac{d\Im(\tau)}{dz}=\frac{d\Im(\tau)}{d\tau}\frac{d\tau(z)}{dz}=\frac{2\tau^*}{H}\exp(-\frac{2z}{H}-\tau^* e^{-2z/H}). \tag{11}$$

It is shown by this simple expression that two factors (the two terms within the exponential function) govern the shape of the weighting function. The first factor leads it to decrease with height manifesting the fact that the density of the absorbing gas decreases with the increase of height. The second factor leads the weighting function to increase with height, which is a result of better transmission of radiation to the space at higher altitudes. The combination of the two competing factors results in a bell-shape weighting function curve peaked at $\frac{H}{2}\ln\tau^*$ (i.e., where $\tau = 1$). Note that when the observing frequency is far from the absorbing line center (τ^* becomes very small), the second factor does not have a measurable contribution, leading to the weighting function peaks at surface. From Eq.(4), the peak of the weighting function is the location where the atmospheric emission contributes the most to the satellite received radiation.

Ideally, we wish the weighting function to be extremely sharp so that the temperature (or water vapor for measuring at H_2O absorption lines) at a distinctive level may be derived from satellite observation at each frequency. However, in reality weighting functions have commonly broad shapes, such as those shown in Fig.2 for NOAA TOVS (TIROS-N Operational Vertical Sounder) channels. To retrieve the vertical profiles of temperature, or water vapor if temperature profile is known, inversion algorithms need to be employed. Physical, statistical and hybrid algorithms have been developed during the past several decades (see, for example, Kidder and Vonder Haar, 1995). A recent development in using sounding channel data is to assimilate the satellite-received radiances directly into numerical weather prediction models (e.g., Eyre et al., 1993), in which case the end products of model fields constitute the retrievals of atmospheric temperature and moisture profiles.

2.3. Microwave Radiation under Cloudy Sky

Scattering by nonprecipitating hydrometeors is generally negligible at low-frequency microwave spectrum (e.g., lower than 40 GHz), since the particle size is much smaller than the wavelength. Thus, Eq.(3) is still applicable for cloudy-sky at microwave spectrum unless there exists measurable precipitation. However, because surface, particularly water surface, is highly reflective in the microwave spectrum, it is convenient to rewrite this equation in the following form

$$I(0,\mu) = \varepsilon_s B(T_s)\exp(-\frac{\tau^*}{\mu}) + \int_0^{\tau^*} B(t)\exp(-\frac{t}{\mu})\frac{dt}{\mu} + (1-\varepsilon_s)\exp(-\frac{\tau^*}{\mu})\int_0^{\tau^*} B(t)\exp(-\frac{\tau^*-t}{\mu})\frac{dt}{\mu}, \quad (12)$$

where ε_s and T_s are the emissivity and temperature of surface. In the microwave spectrum, emissivity of water surface is a strong function of incident angle. The three terms on the right denote, respectively, the transmitted surface emission, emission from the atmosphere (gases and cloud drops), and transmitted portion of the atmospheric emission after it is reflected by the surface. As a crude approximation, one may assume an isothermal atmosphere with temperature being equal to surface temperature, which leads to the following simple form of the equation:

$$I(0,\mu) = B(T_s)[1-\exp(-\frac{2\tau^*}{\mu})(1-\varepsilon_s)]. \quad (13)$$

While simple, this equation makes two important points on microwave remote sensing of clouds. First, since over land the surface emissivity is close to unity for microwave, cloud information (embedded in τ^*) is generally "swallowed" by the strong surface emission. Second, over ocean the surface emissivity is small (on the order of 0.5). The satellite-received radiation largely depends on the atmospheric optical depth. At the lower microwave frequencies used in satellite today (6 – 37 GHz), most of the variations in τ^* under a cloudy sky are due to the change of water vapor and cloud liquid water. In particular, cloud liquid water contributes heavily to the total optical depth at low microwave frequencies except for at the 22.235 GHz water vapor absorption line. The optical depth due to liquid water drops (τ_{LW}) may be expressed as

$$\tau_{LW} \equiv \int_{Z_B}^{Z_T} \beta_{ext}(z)dz \approx \frac{6\pi}{\lambda\rho_w}\mathrm{Im}(-\frac{m^2-1}{m^2+2})L, \quad (14)$$

where Z_B and Z_T are the cloud base and cloud top heights, respectively; $\beta_{ext}(z)$ is volume extinction coefficient of liquid water drops; ρ_w is the density of liquid water; λ is wavelength; m is the complex refractive index of water; and L is liquid water path (vertically integrated liquid water). Clearly, over ocean the satellite received radiation at low microwave frequencies under cloudy conditions is largely determined by the total amount of liquid water in the vertical column.

So far, we did not consider the radiative impact of ice particles. Because the imaginary part of the refractive index of ice is some 3 orders smaller than that of liquid water, the emission from ice particles can generally be ignored at microwave spectrum. However, at high microwave frequencies (>85 GHz), the scattering by ice particles becomes strong enough to significantly alter the satellite-received radiances. An accurate determination of satellite received radiance for the emission-scattering conditions requires solving Eq.(1a) numerically, and the results are not as straightforward as those of the above liquid clouds. However, to get some insight of the ice particles' impact, let's simply assume that the ice cloud layer is above the liquid cloud layer and its scattering acts only to reduce satellite-bound radiation without adding it through multiple scattering. Note that the last assumption only makes sense when the scattering is weak. Therefore, we can modify Eq.(12) as

$$I(0,\mu) = B(T_s)[1-\exp(-\frac{2\tau^*}{\mu})(1-\varepsilon_s)]\exp(-\tau_{IC}), \tag{15}$$

where τ_{IC} is the optical depth due to ice scattering, which may be expressed as follows

$$\tau_{IC} = \frac{24\pi^3}{\lambda^4 \rho_i} |\frac{m^2-1}{m^2+1}|^2 \int_{Z_{iB}}^{Z_{iT}} \int_0^\infty n(r) M_i^2(r) dr dz, \tag{16}$$

for an ice cloud layer located from Z_{iB} to Z_{iT}. In this equation, ρ_i is the density of the ice particles; $M_i(r)$ is the mass of an ice particle with radius of r; $n(r)dr$ denotes the size distribution of the ice particles . To derive this equation, it was assumed that the ice particles are small compared to wavelength so that Rayleigh approximation can be applied. While crude, the above equation captures the essence of the impact of ice particles on satellite-bound high frequency microwave radiation: the reduction due to ice scattering depends on not only the total column integrated ice water amount, but also the detailed size distribution. This feature distinguishes itself from the impact of liquid water on upwelling radiation, in which the cloud's impact only depends on cloud liquid water path regardless the distribution of liquid water drops.

2.3.1 Retrieval of Precipitable Water and Liquid Water Path

In the microwave spectrum, the main absorbing atmospheric gases are oxygen and water vapor. If we use frequencies far from the O_2 absorption line center (~60 GHz) and low enough to ignore ice scattering, the major atmospheric variables that alter the satellite-received radiances are water vapor and cloud liquid water. Let us consider observations of microwave brightness temperatures over ocean at both vertical and horizontal polarizations, denoted as T_{BV} and T_{BH}, respectively. Applying Rayleigh-Jeans approximations and rearranging Eq.(13), we have

$$T_{BV} - T_{BH} = T_s(\varepsilon_V - \varepsilon_H)\exp[-\frac{2}{\mu}(\tau_{O2} + \tau_{WV} + \tau_{LW})], \tag{17}$$

where ε_V and ε_H are the emissivities for vertically- and horizontally-polarized waves at the frequency in question. Note that the total optical depth (τ^*) has been expressed by the summation of optical depths due to O_2 (τ_{O2}), water vapor (τ_{WV}) and liquid water (τ_{LW}). As shown in Eq.(14), τ_{LW} is proportional to liquid water path, L. Similarly, we may express τ_{WV} as a linear function of precipitable water, W, so that

$$k_{WV} W + k_{LW} L = -\frac{\mu}{2}\ln\left[\frac{T_{BV} - T_{BH}}{T_s(\varepsilon_V - \varepsilon_H)\exp(-2\tau_{O2}/\mu)}\right], \tag{18}$$

where k_{WV} and k_{LW} are absorption coefficients for water vapor and liquid water, respectively. If the sea surface temperature and emissivity, and the oxygen optical depth are known, we may derive precipitable water and cloud liquid water path by measuring radiations at horizontally- and vertically-polarized radiations at two frequencies, of which the radiation at one is sensitive to water vapor and the other to liquid water variations (Greenwald et al., 1993). Channel combinations of this type are available in several currently-operating satellite platforms, such as SSM/I, TRMM Microwave Imager (TMI), and Advanced Microwave Scanning Radiometer-EOS (AMSR-E).

2.3.2 Retrieval of Cloud Ice Water

While cloud liquid water retrieval is based on the emission signal from liquid drops at low microwave frequencies, retrieving cloud ice water is based on scattering signal from ice particles at high microwave frequencies. This change brings in additional difficulties to the retrieval problem. First, as shown in Eq.(16), optical depth due to ice depends not only on ice water amount but also the size distribution of ice particles. Second, ice particles are rarely spherical in shape, which complicates the evaluation of scattering. There have been physical algorithms attempting to retrieve both ice water amount and size distribution simultaneously by using multiple channel observations and assuming equal-volume spherical ice particles (e.g., Liu and Curry, 2000; Zhao and Weng, 2002). However, due to the overwhelmingly more number of unknowns than the independent information contents in the problem, these algorithms are far less accurate than those cloud liquid water algorithms described in the previous section.

To perform retrieval for this ill-conditioned problem, statistical methods using *a-priori* knowledge have been developed. One of the methods is based on the Baye's theorem developed by Evans et al. (2002). Baysian inversion methods formally add prior information to that provided by the measurements to obtain a well posed retrieval and corresponding uncertainty estimate. Based on Baye's theorem, the posterior probability density function (PDF) of the cloud ice water path for a given set of measured brightness temperatures ($\vec{T}_B$) can be stated as

$$p_{post}\left(I_i \mid \vec{T}_B\right) = \frac{p\left(\vec{T}_B \mid I_i\right) p_{prior}\left(I_i\right)}{\int p\left(\vec{T}_B \mid I_i\right) p_{prior}\left(I_i\right) dI_i}, \qquad (19)$$

where $p_{prior}(I_i)$ is the prior PDF of cloud ice water path I_i, and $p\left(\vec{T}_B \mid I_i\right)$ is the conditional PDF of the brightness temperature vector given an ice water path I_i. In practice, $p\left(\vec{T}_B \mid I_i\right)$ is commonly assumed to be a multivariate normally distributed PDF with the peak at where radiative transfer model simulated brightness temperatures match those observed, and with the standard deviation proportional to the instrumental and radiative transfer model uncertainties. The *a-priori* PDF, $p_{prior}(I_i)$, is a way to introduce additional information to the algorithm based on prior knowledge. For example, Evans et al. (2002) introduced particle density and size distribution information based on *in situ* aircraft data collected during many field experiments. Seo and Liu (2005) introduced information on the vertical distribution of ice particles based surface-based cloud radar measurements. Once the posterior PDF is determined, the ice water path may be calculated by

$$I = \int I_i p_{post}\left(I_i \mid \vec{T}_B\right) dI_i . \qquad (20a)$$

The uncertainty of the so-retrieved ice water path may be expressed by the variance of the posterior PDF, i.e.,

$$\sigma_I^2 = \int (I_i - I)^2 p_{post}\left(I_i \mid \vec{T}_B\right) dI_i. \qquad (20b)$$

2.3.3 Surface Wind Speed

In the attempt of retrieving atmospheric parameters, we assumed that the surface radiation is known. In reality, the surface emission in microwave frequencies is quite variable, so is the satellite-received radiances, particularly if the frequencies is low (e.g., <20 GHz) where the atmospheric contribution is weak. The variation of surface radiation can be used to estimate soil moisture over land, snow accumulation and sea ice concentration over cold regions, and wind speed over open oceans. The last application is particular important for tropical meteorology because the vast of the tropical region is open ocean.

Over a calm, smooth water surface, the ocean emissivity can be determined by the dielectric property of seawater as a specular surface. However, in the presence of wind, the behavior of the ocean surface emission changes due to the wind-driven roughness and the occurrence of foam; so does the surface emissivity. Therefore, satellite-observed brightness temperature at microwave spectrum changes (increases) with near surface wind speed. This brightness temperature response to surface wind, while currently most relations are still empirical in nature, forms the physical basis for remote sensing of ocean surface wind from passive microwave observations.

2.4. Microwave Radiation under Rainy Conditions

Under precipitating conditions, the scattering by raindrops (and most likely ice particles aloft) becomes so strong that it can no longer be ignored even at microwave spectrum. In this case, the radiative transfer equation of (1a) must be solved numerically with full consideration of multiple scattering. Figure 3 shows the relationship between rainfall rate at surface and upwelling brightness temperature at the top of the atmosphere at 19.35 GHz calculated using a radiative transfer model developed by Liu (1998) for atmospheres with freezing level heights at 1, 3, and 5 km over an ocean surface. At microwave spectrum, Rayleigh-Jeans approximation holds for the Planck's function, so that the brightness temperature is linearly proportional to radiance. In these calculations, no raindrops are assumed to exist above freezing level and vertically constant rainfall rate profiles are assumed below freezing level, although the pattern of rainfall rate profiles has a great variety in actual rainclouds (Liu and Fu, 2001). Because of the polarized nature of water surface emission (reflection), vertically-polarized brightness temperatures are larger than horizontally-polarized ones at 53° viewing angle. As rainfall rate increases, brightness temperature increases at first, but levels off eventually and starts to decrease. The rainfall rate at which the level-off starts depends on the depth of the rain layer; a deeper rain layer causes it to start at a lower rainfall rate. This plot provides the fundamental principle for retrieving rainfall rate over ocean using microwave emission signatures. However, it also spells the challenges facing the algorithm developers. First, until it levels off, the brightness temperature is primarily controlled by the column liquid water path, rather than surface rainfall rate. Thus, the satellite received brightness temperature heavily depends on the depth of the liquid water column, or to be more precise, the detailed liquid water profile. In other words, to retrieve surface rainfall rate, we need to know the vertical distribution of raindrops. Of course, we do not know the raindrop profile when performing retrievals, which adds more unknowns to our inversion problem. Second, the rainfall rate vs. brightness temperature relationship is highly nonlinear, which introduces a so-called (non-uniform) "beam-filling problem" when the rain field is inhomogeneous within a radiometer's field of view (FOV). Most of the modern microwave radiometers have an FOV on the order of 10 km in radius or larger. Within such a large area, rain field is hardly uniform horizontally. The beam-filling problem states that the satellite received radiance depends not only on the satellite FOV-averaged rainfall rate but also on the detailed horizontal distribution of the rain within the FOV. Since the rainfall rate – brightness temperature relations shown in Fig.3 curve downward, it always causes underestimation of the FOV-averaged rainfall rate if not compensating the beam-filling problem in rain retrieval algorithms (Wilheit et al., 1991).

As observing frequency goes higher (but still in the microwave spectrum), radiation originated from the surface and the lower atmosphere will be largely absorbed by the atmosphere and rain layer before reaching to the satellite; the type of surface and the detailed rain structure at the lower atmosphere become less important. Instead, the reduction of upwelling radiation due to the scattering by ice particles aloft becomes the primary signature for retrieving cloud ice amount and rainfall rate. Figure 4 shows the computed brightness temperatures at 85 GHz with 5 km rain layer covered by an ice layer of 1, 3, or 5 km in depth. In the computation, it was assumed that ice particles aloft become bigger and more numerous with the increase of surface rainfall rate. In response to the increase in ice particles aloft, brightness temperatures at the high microwave frequency decrease. On one hand, the results indicate that it is possible to derive cloud ice (including precipitating size ice) amount from observations at high microwave frequencies. However, it should be always understood that the primary signal in this case is ice scattering aloft; therefore, any inference of surface rainfall rate is based on the statistical correlation between ice particles aloft and rain drops reaching to surface. For short-time and small spatial scales, this correlation is not necessarily well established, particularly for clouds that are stratiform in nature. This physical indirectness is the greatest uncertainty in the scattering-based approach in surface rainfall estimation.

2.4.1 Rainfall Retrieval Over Ocean

To better understand the differences of the over-ocean precipitation signatures received by various satellite sensors, we show in Fig. 4 the radiative signatures from a hurricane's clouds, simultaneously observed by several instruments on the Tropical Rainfall Measuring Mission (TRMM) satellite (Liu 2002). The satellite imagery on the left of the figure shows the clouds associated with a hurricane in the South Pacific Ocean. On the right, we show the observed radiative properties at satellite nadir along the line A-B, which crosses the outer cloud band of the hurricane. The parameters shown here include the space-radar-derived rainfall rate distance-height cross-section, the near-surface rainfall rate (also derived from the space-radar), the reflectance at visible wavelength, the thermal infrared brightness temperature and the passive microwave brightness temperatures at 19 and 85 GHz frequencies. Compared to those in the cloud-free area near point B, radiometric properties for rainy areas show the following features: high reflectance in the visible, low brightness temperatures in the thermal infrared, high brightness temperatures at 19 GHz, and low brightness temperatures at 85 GHz. More than half the areas along the line A-B are actually not associated with rain, although clouds in those areas have low infrared brightness temperature and high visible reflectance. It is the microwave brightness temperatures that most closely follow the radar-observed rainfall variation. That is, corresponding to the increase of surface rainfall, the brightness temperature at low microwave frequencies (e.g., 19 GHz) increases, and the brightness temperature at high microwave frequencies (e.g., 85 GHz) decreases. Because of this physical directness between microwave signature and rainfall, microwave satellite measurements have been gaining attention in quantitative remote sensing of precipitation during the past two decades.

As indicated from the above hurricane example, over ocean, both the microwave emission signature due to liquid water drops and the scattering signature due to ice particles can be utilized to infer surface rainfall rate. Several algorithms have been developed to jointly use both emission and scattering signatures, including, for example, Kummerow et al. (1996) who use all microwave channels in a

Bayesian retrieval algorithm, Liu and Curry (1992) who combine 19 and 85 GHz brightness temperatures in a unified function and relate this function to surface rainfall by radiative transfer simulations, and Petty (1994) who uses scattering-based retrieval as a first guess rainfall rate and modifies it gradually to seek the best consistence among brightness temperatures at all channels. The Bayesian algorithm (c.f. section 2.3.2) has been gained tremendous popularity in recent years, particularly due to its flexibility to be adapted for different types of microwave radiometers. This flexibility is important as high temporal resolution of rainfall observation requires combining microwave radiometers from different satellites of various countries to form a constellation of satellites. In a Baysian framework, as long as the instruments are well calibrated and the retrieval algorithms use the same *a-priori* database, the retrievals from different radiometers should not have biases among themselves, which makes them be easily merged.

2.4.2 Rainfall Retrieval Over Land

Over land, the emission signature from raindrops can not be extracted from the upwelling radiometer because the background land surface is overly emissive. A well-used scattering-based algorithm is developed by Ferraro and Marks (1995), in which the brightness temperature at 85 GHz is the primary rainfall indicator. The algorithm takes the form that surface rainfall rate is linearly proportional to the depression of 85 GHz vertically polarized brightness temperature from its clear-sky background value. The clear-sky background brightness temperature is inferred from lower frequency observations that are less sensitive to cloud and precipitation. Recent studies (e.g., Lin and Hou, 2008) indicate that the quality of land rainfall algorithms is far lower than the ocean counterparts, presumably because only scattering signatures are available over land combined with the fact that ice particles aloft have less direct connection with rainfall at the surface. More efforts are certainly needed in exploring new methods to retrieve rainfall over land.

2.5. Reflected Solar Radiation under Overcast Sky

The interaction between solar radiation and clouds are complicated due to the multiple scattering by cloud particles. The intensity of reflected sunlight depends on many factors including surface albedo, cloud optical depth and cloud drop size, just to mention few. In general, the radiative transfer model as given in Eq.(2) needs to be solved numerically to obtain an accurate solution. However, many general properties of reflected sunlight can be explained by the two-stream solution of the radiative transfer model, in which the sunlight is conveniently assumed to be constant at each hemisphere. By further assuming a non-reflective surface, the two-stream solution for reflected sunlight at the top of the atmosphere may written as

$$I^{\uparrow}(0) = I_0 \gamma_\infty \frac{1-\exp(-2\Gamma\tau^*)}{1-\gamma_\infty^2 \exp(-2\Gamma\tau^*)} \quad , \tag{21}$$

where I_0 is the mean solar intensity incident upon the top of the atmosphere, and γ_∞ and Γ are defined as

$$\gamma_\infty = \frac{\sqrt{1-\omega_0 g} - \sqrt{1-\omega_0}}{\sqrt{1-\omega_0 g} + \sqrt{1-\omega_0}} = \frac{1-s}{1+s}, \tag{21a}$$

where

$$s \equiv \left(\frac{1-\omega_0}{1-\omega_0 g}\right)^{\frac{1}{2}} \quad , \tag{21b}$$

and

$$\Gamma = 2\sqrt{1-\omega_0}\,\sqrt{1-\omega_0 g}\ . \tag{21c}$$

While this two-stream solution is crude, it does provide the fundamentals for understanding the cloud reflection of solar radiation, particularly for optically thick uniform clouds. In Eq.(21), the parameter γ_∞, known as the reflection function for a semi-infinite thick cloud, is the major factor for determining the reflected solar radiation. In fact, for a very large τ^*, the upwelling radiative intensity is proportional to γ_∞, which is a function of the similarity parameter, *s*. In Fig.5, the similarity parameter is shown as a function of wavelength (λ) for water clouds having an optical depth of 16 at λ = 0.75 µm, with 5 selected drop size distributions that correspond to effective radius r_e = 4, 8, 12, 16, and 20 µm. Effective radius is defined by the ratio of the 3rd to 2nd momentum of the drop size distribution, *i.e.*, $r_e = \int r^3 n(r)dr / \int r^2 n(r)dr$. At visible band (0.39 to 0.76 µm), the value of *s* is small and independent of the drop size distribution. In contrast, at some near infrared wavelengths the value of *s* varies with drop size distribution as well, with the strongest dependence occurs at 1.6 and 2.2 µm, which indicates that the satellite observed radiances at these wavelengths contain information of both optical depth and particle size. Based on this principle, simultaneous retrieval of both cloud optical depth and effective radius has been proposed as depicted in Fig.6. In the figure, the relationships are shown between reflection functions at 0.75 and 2.16 µm for various values of cloud optical depth and effective radius. Given a pair of observations at these two channels, one can "look up" the values of optical depth and effective radius using this figure. Aircraft observations over a stratocumulus deck are also shown in this figure as circles. In this case, cloud optical depth ranges from 4 to 50, and effective radius ranges from 8 to 32 µm.

2.6 Observations by Cloud and Precipitation Radars

In contrast to radiometers that passively receive energy emitted or reflected by targets, radars actively transmit energy and then receive energy scattered back. The radar-received reflected power P_r can be expressed by

$$P_r = C\frac{|K_w|^2}{R_0^{\ 2}} Z_e \exp[-2\int_0^{R_0} \beta_{ext}(x)dx], \tag{22}$$

where

$$Z_e = \frac{\lambda^4}{\pi^5 |K_w|^2}\int_D n(D)\sigma_{bsc}(D)dD \tag{22a}$$

is called effective radar reflectivity, $K_w=(m^2-1)/(m^2+2)$ and *m* is the refractive index for water, $\sigma_{bsc}(D)$ is the backscatter cross section of a particle with diameter *D*, $\beta_{ext}(x)$ is volume extinction coefficient at distance *x*, R_0 is the distance between radar and the target, *n(D)* is the particle size distribution, λ is radar operating wavelength and *C* is the radar constant. The exponential term in Eq.(22) is due to the attenuation by hydrometeors and atmospheric gases between radar and targets, and often ignored (set to 1) in practice. If the targets are small water drops, their backscatter cross sections are approximately proportional to D^6, and the effective radar reflectivity may be simplified as

$$Z_e = Z \equiv \int_D n(D) D^6 dD , \tag{22b}$$

and Z is referred to as the radar reflectivity. In addition, for ice particle,

$$Z_e = \frac{|K_w|^2}{|K_i|^2} Z , \tag{22c}$$

where K_i has the same form as K_w but m is the refractive index for ice.

Effective radar reflectivity is often used to estimate precipitation rate R and cloud/precipitation water content W_c using the so-called Z-R (or Z-W_c) relation. Since Z is proportional to D^6 while W_c (R) is proportional to D^3 ($\sim D^{3.5}$ because terminal velocity proportional to $\sim D^{0.5}$), obviously, the Z-R or Z-W_c relationship varies with the particle size distribution, which is the major factor for the uncertainties in radar-derived rainfall rate. Constraining the relations using observations at dual wavelength is an active research topic.

2.7 Ocean Surface Wind Measurement by Scatterometers

Selecting a relatively low operating frequency (long wavelength), the radar-received backscattered energy from atmospheric targets becomes less significant, and is dominated by reflection from surface. Over the ocean, wind stress generates capillary-gravity waves that backscatter radar-transmitted energy by the mechanism of Bragg-scattering (and to a lesser extent, by "sea spikes", e.g., Phillips, 1988). The backscatter cross section, σ_0, depends on near-surface wind speed, the azimuth angle between the incident radiation and the wind vector, radar beam zenith angle, wave polarization, etc. Figure 7 shows the relation between σ_0 and wind direction for wind speeds of 5 to 30 m/s (Naderi et al., 1991). It is seen that the backscatter increases with wind speed and is greater at upwind or downwind (0° or 180° relative direction in the figure) than at crosswind directions. Due to the no-uniqueness in the relation between σ_0 and wind speed and/or direction, a single scalar scatterometer measurement is insufficient to retrieve wind speed or direction. A satellite scatterometer is designed to measure the same ocean surface spot from several different azimuth angles in order to determine wind speed and direction.

3. Applications of Satellite Observations in the Tropics

3.1 Hurricane Studies

In the tropics, the most distinctive weather phenomenon is hurricane. Note that for simplicity, here we use the word "hurricane" for tropical cyclones at all developing stages including tropical depression, tropical storm, and hurricane 1 to 5 (in the Saffir-Simpson scale) in the NOAA classification. The importance of satellite observation for hurricane studies may be readily seen from Fig. 8, in which the number of "observed" Atlantic hurricanes (named storms) and the percent of those struck land are shown for the period of 1900 to 2006 (Landsea, 2007). There is a clear jump of the percentage of hurricanes that struck land around 1965 when satellite observations started to be used for hurricane detection. Before then, hurricanes had been reported only by land and ship based observers; thus significant number of storms that dissipated before reaching land may have been uncounted for. The multisensory observations from satellite in recent years have not only improved our ability of hurricane detection, but also enable us to have a better assessment of its intensity and a better understanding of its physical processes.

3.1.1 Intensity

The idea of estimating the intensity of hurricanes from satellite data has been developed during the early years in the history of satellite meteorology. Particularly worth mentioning is the technique developed by Dvorak (1975; 1984) during the 1970's and still been routinely used today, in which the intensity of a hurricane is determined by its appearance in visible and/or infrared imageries. Utilizing a satellite imagery of a tropical cyclone, the method starts matching the image versus a number of possible pattern types: curved band pattern, shear pattern, eye pattern, central dense overcast pattern, embedded center pattern or central cold cover pattern. If an infrared satellite imagery is available for eye patterns, the method also utilizes the difference between the temperature of the warm eye and the surrounding cold cloud tops. The larger the difference, the more intense the tropical cyclone is estimated to be. From this one gets a T-number, which is then empirically translated to the maximum wind and minimum central pressure of the hurricane. Shown in Fig.9 are examples of the tropical cyclone patterns and their associated T-numbers. While being purely empirical and subject to certain ambiguity depending on observer's experience, the Dvorak method does provide a relatively consistent estimate of hurricane's intensity in the absence of other observations, such as aircraft reconnaissance. To avoid subjectivity, a computer-based version, called "Objective Dvorak Technique", has been developed by Velden et al. (1998). Detailed information about this more objective method is available at http://cimss.ssec.wisc.edu/tropic/research/products/dvorak/odt.html).

A distinctive feature of a hurricane is its warm core. Another idea to infer hurricane intensity from satellite observations is therefore to measure the temperature anomaly of the warm core. Under the assumption of hydrostatic balance in the core region, the magnitude of the warm temperature anomaly in the warm core may be physically related to the pressure anomaly (i.e., the hurricane intensity) at the hurricane center (e.g., Kidder, 1979). As described in section 2.2.2, atmospheric temperature profile may be retrieved by measuring the upwelling radiances at multiple channels near CO_2 and/or O_2 absorption lines. For hurricane intensity estimation, observations at the microwave frequencies are particularly advantageous because they are less sensitive to the contamination by non-precipitating clouds. A cross section of the temperature anomaly for Hurricane Floyd on 11 September 1999 retrieved from the Advance Microwave Sounding Unit (AMSU) data is shown in Fig. 10, indicating the warm anomaly aloft is about 7 K in the storm's core in contrast to the cooling in the low level rainbands of about -7 K. An operational algorithm has been developed by Velden et al. (1991) and Brueske and Velden (2003) using data from Microwave Sounding Unit (MSU) or AMSU on NOAA polar-orbiting satellites. In this algorithm, the minimum sea level pressure and maximum sustained surface wind (from best track data) are linearly regressed against MSU-derived temperature anomaly at 250-mb level or brightness temperature anomaly at AMSU-A 54.94-GHz channel (weighting function peaks near 250 mb). Compared to the intensity determination by pattern (Dvorak technique), the temperature anomaly based approach is more objective although the intensity-warm anomaly relation is still empirically derived. However, one significant shortcoming of this approach arises from the fact that the spatial resolution of AMSU O_2 sounding channels is too coarse (~50 km at nadir and ~150 km at swath edge). So that, more often than not, it is not possible to obtain an AMSU observation entirely from the eye, but rather pixels near the storm center often cover part of the eyewall area. This mix of signals from both core and eyewall makes it difficult to estimate the actual warm temperature anomaly in the storm core.

A less mature but rather interesting idea to estimate hurricane intensity has been proposed by Wong and Emanuel (2007) based on the assumption that a hurricane is a vortex in gradient and hydrostatic balance, so that its peak wind speed, V_m, may be expressed by

$$V_m^2 \approx \frac{T_s - T_0}{T_0}\Delta h^*, \tag{23}$$

where T_s and T_0 are sea surface temperature (SST) and cloud-top temperature, Δh^* is the change in saturation moist static energy ($\equiv c_pT+gz+L_vq$, where T, z, and q refer to the temperature, height from the surface, and specific humidity, respectively. c_p is the isobaric specific heat capacity of dry air, g is gravity, and L_v is the specific latent heat of vaporization.) at cloud-top level, from the eyewall h_e^* to outer region h_0^*. Since water vapor density at cloud-top level (~15 km) is negligible, the saturation moist static energy is primarily determined by the cloud-top temperature and height. Luo et al. (2008) examined this technique using coincident observations of infrared cloud-top temperature from Moderate Resolution Imaging Spectroradiometer (MODIS) and cloud-top height from CloudSat cloud radar, in addition to weekly SST. Figure 11 shows one example of such coincident observations when CloudSat passes over Hurricane Ileana's eye on 23 August 2006 while MODIS covers almost the entire hurricane cloud system. As evident in Fig. 11, cloud-top height in the outer region tends to be irregular, which makes it difficult to assign a unique value for h_0^*. For this reason, an alternative method is also attempted to estimate h_0^* based on the idea that the saturation moist static energy at the tops of the outer convective clouds will be approximately equal to the actual moist static energy of undisturbed air in the boundary layer. Therefore, h_0^* is so estimated by assuming that the surface air temperature is equal to the SST, with a relative humidity (RH) of 80%. Figure 12 shows the comparison of the satellite-estimated against best track maximum sustained wind for 9 hurricanes coincidently observed by CloudSat and MODIS during the first half-year of CloudSat operation. It is evident that the method described by Eq.(23) clearly has significant skills in estimating hurricane intensity although it seems to produce overestimation when the storm intensity is low.

3.1.2 Structure

Multispectral and multiplatform satellite observation provides a powerful way to view the structure of a hurricane from different perspectives. In Figs.13 and 14, we would like to show the structural differences between Hurricane Dennis on 10 July 2005 and Hurricane Rita on 21 September 2005, both in the Gulf of Mexico. The visible images from geostationary satellite are shown in Fig.13, in which both storms show well-defined storm centers and solid circular cloud shield filled the inner region of the storm (~250 km in radius). Both storms were moving to northwest at the time. From AMSR-E and AMSU-B microwave observations, Dense ice index (Liu et al., 1994), ice water path (Seo and Liu, 2005), liquid water path (Liu and Curry, 1993) and rainfall rate (Liu and Curry, 1992) are computed and their distributions are shown in Fig.14. The difference between dense ice index and ice water path is that the former only counts for dense ice particles such as graupels and hails while the latter also counts for ice particles with lesser density. Liquid water path is not retrievable over land areas. While the distributions of the dense ice index and rainfall are quite similar, they are different from those of ice and liquid water paths, manifesting that cloud and precipitating water are not entirely collocated in the vertical. In the figure also shown is the storm motion (black arrow) and vertical shear (between 850 and 200 mb) directions (red arrow). Very strong vertical shear was observed for Dennis case on 10 July 2005 in the direction approximately normal to the direction of the storm's motion, while there was very little vertical shear for the Rita case on 21 September 2005. Compared to Dennis, the hydrometeors are much more axisymmetric about its center, particularly in the inner region.

This preferential distribution of rainfall has been reported by Rogers et al. (2003) using a mesoscale model and by Chen et al. (2006) using TRMM radar data. Figure 15 shows the asymmetry of the rainfall distribution derived from 3-year TRMM rainfall data over the globe (Chen et al., 2006). It is shown that regardless the storms' intensity the rainfall preferentially falls in the downshear-left quadrant, consistent with the case study of Dennis in Fig.14. They further found that this asymmetry can also be seen in the composite of vertical distributions of radar reflectivity (Fig. 16), which is a proxy of the amount of precipitating (liquid and/or ice) particles. According to their analysis, the vertical wind shear is a dominant factor for the rainfall asymmetry when shear is greater than 5 m s^{-1}. The storm motion-relative rainfall asymmetry in the outer rainband region is comparable to that of shear-relative when the shear is smaller than 5 m s^{-1}.

3.1.3 Water Balance

Using a suite of satellite and ground-based conventional data, Liu et al. (1994; 1995) attempted to estimate the water balance in a Pacific typhoon (Typhoon Nina, 18-28 November 1987). The water balance in a radial area was expressed by the following equation:

$$P + \frac{\partial C}{\partial t} + \frac{\partial V}{\partial t} = HT + E\,, \tag{24}$$

where P represents the removal of water by precipitation, $\partial C/\partial t$ the storage of condensed water (liquid and ice), $\partial V/\partial t$ the storage of water vapor, HT the net horizontal transport of water vapor into the area, and E the rate of evaporation from the ocean. Further, they divided the storm's life time into formation, developing, mature and decaying stages. Table 1 shows the magnitude of the five terms in Eq.(24) for three (in 1°, 2.5°, or 4° latitude/longitude radius) radial areas.

Table 1. Water Balance in 1°, 2.5° and 4° Radial Areas for Typhoon Nina (from Liu et al., 1994)

Stage of the typhoon	P (mm h^{-1})	$\partial C/\partial t$ (mm h^{-1})	$\partial V/\partial t$ (mm h^{-1})	E (mm h^{-1})	HT (mm h^{-1})	% of HT to total source	Ratio of E to HT
1° Radial Area							
Formation	3.28	0.04	-0.05	0.72	2.55	78	0.28
Developing	9.38	0.02	0.01	0.72	8.69	92	0.08
Mature	11.28	0.04	-0.14	0.72	10.36	94	0.07
Decaying	6.44	0.01	0.06	0.72	5.79	89	0.12
2.5°Radial Area							
Formation	1.32	0.02	0.06	0.51	0.87	66	0.59
Developing	4.56	-0.003	0.03	0.51	4.08	89	0.13
Mature	4.15	0.02	-0.09	0.51	3.57	88	0.14
Decaying	3.25	-0.01	0.05	0.51	2.78	84	0.18
4° Radial Area							
Formation	1.12	0.01	0.03	0.29	0.87	75	0.33
Developing	2.25	-0.01	-0.04	0.29	1.91	87	0.15
Mature	1.88	-0.01	-0.09	0.29	1.49	84	0.19
Decaying	1.87	-0.02	-0.36	0.29	1.20	81	0.24

The results in Table 1 indicate that values of storage terms ($\partial C/\partial t$ and $\partial V/\partial t$) are generally two orders of magnitude smaller than rainout term (P), except for $\partial V/\partial t$ during the decaying stage in the 4° radial area. In addition, surface evaporation in the vicinity of the storm is also about one order smaller than rainout term, which makes that the major water balance within 4° radial area of the storm is dominated by the removal by precipitation and convergence of water vapor from surroundings. Except for at the formation stage, horizontal transport counts more than 80% of the total water removed by precipitation within the radial area of 4°.

Liu et al. (1995) examined the water vapor supply for several cloud clusters in the same region as Nina before it was developed to a named storm. Satellite images show that several cloud clusters developed in the same area several days before the tropical cyclogenesis (Fig. 17). By examining the sea surface evaporation and precipitation in the cloud clusters, it was found that the precipitation exceeded evaporation by several times in the precipitating areas of the clusters, indicating that local evaporation alone could not supply enough water vapor, and horizontal transfer of water vapor from surrounding areas is required for the storm to continue to develop. Surface wind field (Fig. 18) indicated that there was a constant increase of cyclonic wind in the area of the cluster (F) that finally led to the tropical cyclogenesis, while no apparent increase of cyclonic wind was found in the other cloud clusters. In addition, water vapor amount in Cluster F did not decrease for several days before the disturbance was upgraded to a tropical storm, while it was found to decrease after the mature stage for the other cloud clusters that did not evolve into tropical storms (Fig.19). From consideration of the water vapor balance, Liu et al. (1995) interpreted the cyclogenesis as a transition from an unbalanced cluster to a balanced cluster. Horizontal transfer of water vapor in a water vapor-unbalanced cloud cluster is not large enough to overcome the deficit caused by precipitation over evaporation. The shortage of water vapor in the unbalanced cluster results in a short-lived cloud cluster. When the sum of evaporation and horizontal transfer can provide enough water vapor supply to balance the removal by precipitation (balanced cluster), the precipitation does not “dry up” the atmosphere. This is the necessary condition for the cyclogenesis. The increase in horizontal transfer of water vapor is found to be associated with the increase of the surface cyclonic wind.

The surface cyclonic wind as a strong indicator for cyclogenesis is further recognized by Sharp et al. (2002), who developed an algorithm for “early” detection of tropical storms using satellite scatterometer surface wind-derived vorticity. Figure 20 shows several examples of storms in the 1999 Atlantic hurricane season, in which the scatterometer wind indicated cyclonic circulations, thus potential cyclogenesis, one to three days before the official declaration of the formation of tropical storms. Therefore, scatterometer data are potentially useful for improving current tropical storm detection techniques.

3.2 Precipitation Measurements

Latent heat release from tropical convections is the driving force of the global atmospheric circulation. Thus, observing the magnitude, spatial distribution and temporal variability of clouds and precipitation in the tropics has long since been one of the primary goals of atmospheric research. Since vast of the tropical region is covered by ocean, surface-based rain gauge measurements of rainfall can merely capture a small fraction of the whole picture of the tropical precipitation. Shown in Fig.21 are the current surface-based rain gauges over the globe. It is seen that over 60% of the rain gauges are located between 30°N and 60°N, and over the tropical oceanic regions there are only sparsely spaced stations over atolls. Therefore, measurements of rainfall in the tropics will naturally have to rely on satellite observations. One milestone in satellite rainfall measurements is the TRMM mission. Launched in 1997, the primary sensors for rainfall measurements on the TRMM satellite are a precipitation radar and a multi-frequency microwave radiometer (Kummerow et al., 2000). In the following subsections, many results are primarily derived from these two sensors.

3.2.1 Rainfall Climatology

With the goal of developing of a more complete understanding of the spatial and temporal patterns of global precipitation, the Global Precipitation Climatology Project (GPCP) has been established to collect and merge precipitation data from rain gauges and a variety of satellite retrievals (Adler et al., 2003), which are widely used by scientists in model validation and climate diagnostic studies. Figure 22 shows the rainfall rate maps in the tropics averaged over 10 years (1998-2007) generated from GPCP merged satellite and rain gauge data (monthly, 2.5° resolution). While GPCP data at high latitudes (45° polarward) still have large uncertainties, its rainfall estimates in the tropics are of high confidence largely because of the addition of TRMM data, which have been inter-compared with multiple ground-based measurements. Rain bands are shown in the inter-tropical convergence zone (ITCZ) and along the mid-latitude storm tracks. The zonally averaged rainfall rate (Fig.23) in the ITCZ latitudes is about 6 mm day^{-1}, translating to over 2 m of rainfall annually, while in the subtropics the zonal mean is about 2 mm day^{-1}. Over the vast oceanic regions, the main input for this rain product is satellite retrievals.

The seasonal variation of the rainfall distribution over the tropics is presented in Figs.24 and 25. Rain maximums are very clear over the Indian monsoonal region during June-July-August and over the Amazon rainforest region during December-January-February. In the zonally averaged rain map (Fig.25), the location of the rainfall maximum varies from ~5°S during Northern hemisphere winter to ~10°N during Northern hemisphere summer, following the movement of ITCZ, with its magnitude reaching the peak at ~ 9 mm day^{-1} around July to August. The rainfall minimum occurs in the subtropics during winter seasons with a magnitude of lower than 1 mm day^{-1}.

The dominant signature of interannual variability in tropical precipitation is the different rainfall distributions between El Niño and La Niña years. Launch in November 1997, the TRMM satellite had the opportunity to observe the El Niño event during Northern hemisphere winter of 1997-98 and the La Niña event after it (Northern hemisphere winter of 1998-1999). Figure 26 shows the TRMM observed mean rainfall distribution during the two months of January and February 1998 (El Niño), and the difference in the same two-month mean rainfall between 1999 and 1998 (subtracting the mean of 1998 from that of 1999) (Murtugudde et al., 2000). SST anomaly (not shown) depicts the peak of the 1997-1998 El Niño at the start of the year and the entry into the 1998-1999 La Niña by the end of the year. The corresponding changes in the tropical precipitation field are characterized by a dipole pattern: a belt with reduced precipitation from the eastern equatorial Pacific Ocean to the Central part, and an area with increased precipitation over the western Pacific Ocean and the southwestern Pacific Ocean.

3.2.2 Horizontal and Vertical Variability of Rain Field

Precipitation radar (PR) on TRMM satellite provides a unique opportunity to study the vertical and horizontal structures of clouds and precipitation in the tropics. By examining the existence of bright-band and horizontal variation, PR is able to determine not only the intensity of precipitation, but the precipitation type (convective, stratiform, warm rain, etc.) as well, because stratiform rain echoes are commonly associated with bright-band near the melting level and convective rain echoes have greater horizontal variability. Additionally, warm rain events can be determined by echo-top height (lower than freezing level height). The number ratio of a particular rain type to total observations (i.e., area fraction) is calculated for each 1°x1° box during the entire year of 1998 and its horizontal distribution is shown in Fig. 27. The zonally averaged values of the number ratio are shown in Fig. 28. On zonally average, stratiform has a ratio of ~3-4% while the ratios for deep and warm convections do not exceed 1%. Stratiform rains are the dominant precipitation type in terms of area coverage, and have three maxima in the zonal frequency distribution: one near the equator and two in the mid latitudes. Deep convections occur most frequently in the ITCZ. Unlike stratiform rainfalls, the occurrence frequency of deep convections does not have peaks in the mid latitudes. Warm convections have higher frequencies in the subtropical oceanic regions in both hemispheres, particularly in north Pacific, southeast Pacific, south Indian, and southwest Atlantic Oceans. The abundance of warm shallow precipitation near Hawaii is clearly shown in Fig. 27c.

The vertical distributions of rain profiles show distinct differences between convective and stratiform rain events, and between rain events over land and ocean, as shown in Fig.29, which are derived from PR observations between 15°S and 15°N during the year of 1998. Over ocean (Figs.29c and d), for stratiform rains, the bright-band near 4.5 km separates two totally different variation regimes. Below the bright band, rainfall rates are almost constant with only a slight decrease toward the surface, implying that raindrops experience no further growth or may undergo slight evaporation. Above the bright-band, a sharp downward growth of rainfall rates occurs within a layer 1 km deep, followed by a slower growth above. It is explained that while falling from above, ice particles grow mainly by vapor deposition at high level, which is the slowest microphysical growth mode. While they are near freezing level, the ice particles may grow by deposition, riming, and more importantly, aggregation. The last process increases the particle size dramatically, which in turn sharply increases radar reflectivity. The bright-band itself is a result of melting process, that does not reflect the change in mass flux. The mean profiles for deep convections display different characteristics from those of stratiform profiles. The maximum of rainfall rate appears at a much lower level (2-3 km) than the height of the stratiform bright band. A tendency of decreasing rainfall rate toward the surface can be seen below the maximum rainfall rate level, especially for heavier precipitations, which indicates the possibility of raindrop evaporation. Above the maximum rainfall rate level, we can generally divide the vertical distribution into three regimes: (1) a slow drop off from the maximum level to freezing level (~5 km), (2) a rapid drop off in the layer of 1 to 2 km in depth above freezing level, and (3) another very slow drop off above. While the pattern above freezing level is quite similar to those in stratiform profiles, the additional growth (toward surface) of rainfall below freezing level is unique to the convective profiles, which may be an indication of growth by coalescence. The mean profiles of rainfall over land are shown in Figs. 29a and b. The profiles for stratiform rainfalls are very similar to those over ocean. However, the deep convective profiles have significant difference from those over oceans. The height of the maximum rainfall rate is at 3 – 4 km, higher than that in the over ocean profiles. As a result, the layer where it may correspond to growth by coalescence becomes very shallow. For the same surface rainfall rate, the growth layer above freezing level is deeper than that in the ocean cases, presumably due to stronger updraft for over land convections. Below the maximum rainfall rate level, the drop off of rainfall rate toward surface becomes more evident than those over ocean cases. This may be an indication of more server evaporation over land areas. It is interesting to see that there is no visible difference between over ocean and over land stratiform rains. The most significant difference between deep convections over ocean and over land appears to be that land rains have a deeper rain layer.

The characteristic horizontal variability of rain field within a scale similar to the field-of-view of satellite microwave sensors (25 km x 25 km) has been investigated by Varma and Liu (2006) over the global tropics using three years (Dec. 1997 – Nov. 2000) of TRMM PR data. In the data analysis, the 25 km x 25 km rainy "pixels" are categorized by rain type (convective or stratiform), rain intensity (light, moderate or heavy), surface type (land or ocean) and latitudinal location (tropics or extratropics). Two attributes are used to characterize the small-scale rainfall rate variability: fractional rain cover, or FRC, and conditional rainfall rate probability density function, or PDF. It is found that FRC is closely related to the areal averaged rainfall rate, R_{av}, and the greatest cause in varying the FRC – R_{av} relation is rain type (Fig.30). Given the same R_{av}, FRC for stratiform rain is larger than FRC for convective rain. FRC increases with R_{av} to 100% at R_{av} ~3 mm h^{-1} for stratiform rain and ~10 mm h^{-1} for convective rain. For a convective rain event, even the averaged rainfall rate over a 25 km by 25 km area is as high as 5 mm h^{-1}, 30% of that area may still be rain-free. This figure vividly illustrates how variable the horizontal rain field could be. Figure 31 shows the conditional PDFs of rainfall rates derived from the 3 years of TRMM PR data for 3 categories of field-of-view averaged rain intensities: light rain (< 2.5 mm h^{-1}), moderate rain (2.5 – 10 mm h^{-1}) and heavy rain (> 10 mm h^{-1}). Convective rain PDFs are broader over all rain intensity ranges, but particularly evident when the field-of-view averaged rainfall rates are light or moderate. For convective rain over land (lines with hollow symbols), the PDFs of rainfall rates do not show a difference between the tropics and extratropics regardless of the intensity of rain. However, the land-ocean difference in convective rainfall rate PDFs is particularly evident for the light and heavy rain categories. For stratiform rain in the light rain category, the four PDFs of rainfall rate do not show much difference among them. Although the difference among them increases as rainfall intensity increases, the land-ocean difference continues to dominate the variability of the PDFs. The influence of latitudinal location on the pattern of rainfall rate PDFs is greater for heavy rain than for light rain.

3.2.3 Diurnal Variations

The TRMM satellite is non-sunsynchronous, which allows observing the Earth at different local time of the day. The diurnal variation of rainfall then can be studied by composite data during a sufficient long period. Figures 32 and 33 show results from Takayabu and Kimoto (2008) who computed the amplitude and phase of the first harmonic component of the diurnal cycle using 5-year (1998-2002) TRMM PR and TMI rainfall estimates during Northern Hemisphere summer months (June through August). It is shown that the diurnal cycle has higher amplitudes (Fig.32) over land and coastal regions than over open oceans, while rainfall peaks in the afternoon over land and early morning over oceans (Fig.33). The difference in the diurnal cycle phases manifests the different mechanisms that enhance precipitations over land and ocean. That is, over land atmospheric instability resulted from solar heating of surface is the dominant mechanism for rainfall enhancement, while over ocean the main factor becomes the radiative cooling at cloud-top.

To gain further insight of the rainfall diurnal cycle, Nesbitt and Zipser (2003) performed analysis of TRMM observed rain events by dividing them into three kinds of "precipitation features" (PF) based on PR reflectivity and TMI 85 GHz PCT (polarization correct temperature, Spencer et al., 1989). The PCT, defined as PCT=$1.818T_{BV}$-$0.818T_{BH}$, is a measure of the amount of precipitating ice in the clouds (A smaller PCT indicates a greater amount of precipitating ice). A PF in their definition is a contiguous area larger than 75 km^2 with either PR near-surface reflectivity ≥20dBZ or 85 GHz PCT≤250K. The three kinds of PFs they defined are: (1) PF without ice scattering (no pixels containing 85 GHz PCT≤250K), likely "warm rain" features too shallow or those too small to contain significant ice scattering at TMI resolution; (2) PF with ice scattering (at least one pixel with 85 GHz PCT≤250K), features with significant precipitating ice aloft, but not large enough or intense enough to meet the MCS (Mesoscale Convective System) category; (3) PF with an MCS (PF contained at least 2000 km^2 of contiguous area with 85 GHz PCT≤250K and 185 km^2 ≤225 K). Figures 34 and 35 show the following several characteristics of the diurnal cycles PFs by type over land (Fig.34) and ocean (Fig.35) areas: (a) volumetric rainfall (sums of every PF's average rainfall rate times its area), (b) number of PFs by type, (c) median area of each feature type, and (d) mean conditional rainfall rate for each type of PFs. Over land, all three PF types modulate the diurnal cycle. There is a sharp early afternoon peak in overland rainfall (around 1500 LT), enhanced by a peak in rainfall from PFs with and without ice scattering that both peak at this time. PFs with and without ice scattering over land diurnally vary in number (Fig.34b) and mean conditional rainfall rate (Fig.34d) in a similar fashion as the total volumetric rainfall (all peaking in the early afternoon). PFs with overland MCSs, however, have distinctly different patterns in diurnal variation. While their mean conditional rainfall rates remain nearly constant throughout their daily cycle, the number of MCSs peaks in the evening (1700-1900LT) and their median area has the maximum in the morning (0900LT). As a result, the total rainfall contributed by MCSs has shallow peak near 0100 LT (Fig.34a), which is different from the afternoon peak of overall rainfall.

Over ocean (Fig.35), the magnitude of the diurnal cycle is considerably less than that over land. The total rainfall reaches a maximum near sunrise (0600 LT). Unlike land areas, all oceanic PF types peak in rainfall contribution in the early morning hours. Since the mean conditional rainfall rates and the median areas of the three PFs have little variation during the day, the diurnal cycle of rainfall over the ocean seems almost completely to be due to an increase of the number systems, not the rainfall rates contained in them.

3.2.4 Latent Heating

Besides driving the global hydrological cycle, a consequential effect of precipitation is the release of latent heat to the atmosphere due to the phase change of water, which serves as the primary fuel for the tropical convective heat engine. One of the primary goals of TRMM mission is to derive latent heating distributions from satellite observations. A summary of the progress in developing latent heating retrieval algorithms was presented by Tao et al. (2006). It is noted that the methodology in retrieving latent heating is fundamentally different from that in retrieving rain. In the retrieval of rainfall, the targets to be detected (raindrops) are actually "objects" that interact with electromagnetic waves, so that the "amount" of the targets can be somehow related to the satellite received energy through either radiative transfer model [Eq. (1)] or radar equation [Eq.(22)]. However, latent heating is a result of phase change of water and there is no established mathematical equation to connect it to satellite received energy. Therefore, current retrieval of latent heating has to heavily rely on cloud resolving models (Tao et al., 2006). The basic idea of latent heating retrieval algorithms is as follows. First, cloud resolving model simulations are performed for various surface and atmospheric conditions, which should be as diverse as possible. From the simulation results, a lookup table is then created linking latent heating profiles to certain "cloud and precipitation features". Depending on algorithm configuration, the "cloud and precipitation features" can be hydrometeor profiles, surface precipitation rate and types (e.g., stratiform, convective, warm rain, etc.), cloud top height, bright-band height, etc. In performing the retrieval, latent heating profiles are found from the lookup table using satellite-measured "cloud and precipitation features". Because of the heavy reliance on cloud resolving model results, how well the cloud resolving model performs and how representative the simulated cases in the lookup table are will largely dictate the performance of the retrieval algorithm. Therefore, future research to improve the retrieval method is certainly needed.

An example of the latent heating retrieval for an Atlantic tropical cyclone (Hurricane Bonnie) is shown in Fig.36 (Tao et al., 2006). The structure of the hurricane eye and convective rain spiral bands are properly captured. Precipitation mass appears at high altitudes in the presence of deep convection, such as around 200 km along the satellite's nadir track (middle panel). Widespread weaker rainfall rates are found between convective cells. In the weaker rain areas, rainfall rates are mostly concentrated in the middle to lower troposphere where stratiform conditions are prevalent. Deep latent heating is associated with the strongest convective cells. Peaks of maximum latent heating vary with different conditions and their altitudes are generally at or below 5 km. Evaporative cooling occurs near surface and in the lower troposphere in the stratiform regions.

3.3 Clouds and Surface Energy Fluxes

Figure 37 is a schematic of the global energy budget (Trenberth et al., 2009), which shows about 50% of the solar energy reached to top-of-atmosphere (TOA) will be absorbed by the surface. While about 40% of the surface energy gain is returned to the atmosphere in the form of (net) longwave radiation, the rest will be transferred by sensible and latent heat fluxes. Much of the upward heat transfer occurs in the tropical ocean. The net downward heat flux at ocean surface, F_{net}, may be expressed by

$$F_{net} = F_{sw} - (F_{lw} + F_{sh} + F_{lh}), \tag{25}$$

where F_{sw}, F_{lw}, F_{sh}, and F_{lh} are the net downward shortwave radiation, net upward longwave radiation, sensible heat flux and evaporative latent flux, respectively. Unlike at the TOA where radiative fluxes can be directly measured by sensors on satellite (e.g., ERBE: Earth Radiation Budget Experiment, and CERES: Clouds and the Earth's Radiant Energy System), deriving surface energy fluxes requires much deliberated retrieval procedures, and their current best estimates contain greater uncertainties (Trenberth et al., 2009). One of the complications in surface energy fluxes' estimation is due to clouds, which is particularly true for the radiative fluxes. In this section, we first introduce research results of satellite data derived cloud properties, followed by studies on surface heat fluxes in the tropics.

3.3.1 Cloud Properties

The International Satellite Cloud Climatology Project (ISCCP, Rossow and Schiffer, 1991; http://isccp.giss.nasa.gov/products/isccpDsets.html) dataset is probably the most complete and comprehensive satellite dataset on cloud properties ever produced so far. Derived from multi-platform and multispectral (primarily visible and infrared though) satellite data with state-of-the-art radiative transfer models, ISCCP dataset contains global cloud macro- and microphysical, and radiative properties at ~30 km resolution every 3 hours since 1983.

In Fig.38, the global distribution and its zonally average of cloud amount (%) are shown, which are derived from visible and infrared satellite data in the duration of July 1983 through June 2008. High cloud amounts are seen in the mid latitude storm tracks flowed by the belt of ITCZ. Low cloud amounts are found in the subtropical regions, with the lowest value at the North African desert. The highest zonally averaged cloud amount occurs around 55°S with a peak value of ~90%. The global mean value of cloud amount is ~65%. The long-term variation (Fig.39) of global mean cloud amount is small (±2%), although there is a clear annual cycle. It also suggests a multi-decadal oscillation with a peak near 1987 and a valley near 2000. For cloud optical depth for shortwave (Figs.40 and 41), the zonally-averaged value in the tropics and subtropics (±30° latitudes) is about 2~3, much lower than those in the high latitudes. It should be mentioned here that the optical depth retrievals in the polar regions are not as reliable because of the low solar elevation angle. The global averaged value for cloud optical depth is ~4. Beside annual cycle, it does not seems to have a clear pattern of long-term variation during the 30+ years record, although there seems to be a increasing trend in the most recent decade.

One cloud microphysical property, the effective radius of cloud drops, has been investigated by many investigators using satellite. In Fig.42, we show the results by Han et al. (1994), who derived the monthly mean values of effective radius from 1987 ISCCP data for the ±40° latitudinal regions. The most striking feature from these retrievals is the contrast of particle size between land and ocean environment; cloud drops are categorically smaller over land than over ocean. Generally speaking, the mean effective radii are smaller than 8 μm over land while they are larger than 10 μm over ocean. The difference may be explained by the difference in the number concentrations of cloud condensation nuclei between land and ocean areas.

A bi-spectral (visible and thermal infrared) analysis of cloud properties can classify observed clouds into different categories. Figure 43 is such a cloud classification scheme proposed by Rossow and Schiffer (1991). In the diagram, one axis is cloud optical thickness (i.e., optical depth) derived from visible measurements, and the other axis is the cloud top pressure measured by thermal infrared sensors. Although the boundaries are drawn in a slightly arbitrary fashion, the position of an observed cloud in this diagram tells the type of the cloud, e.g., deep convection, stratus, etc. A certain pattern of cloud occurrence frequency in the 2-dimensional histogram can be linked to a synoptic regime (Jakob and Tselioudis, 2003). In the following, we show a study by Rossow et al. (2005), in which they described the typical cloud climatology in the tropics by several synoptic regimes determined by the above bi-spectral analysis. In the analysis, they used 21.5 years (1983-2004) 3-hourly ISCCP data covering the tropical area of ±15° latitudes. Using the K-mean clustering technique (Anderberg, 1973), they looked for distinctive patterns in the joint frequency distributions of the cloud top pressure and optical depth values from individual satellite image pixels (5 km in size) occurring within 2.5° regions. Based on their analysis, they found that there are 6 distinctive patterns of the joint frequency distributions, which they referred to as "weather states" (WSs). The averaged histograms for the 6 weather states are shown in Fig.44, along with their relative frequency of occurrence (RFO), in order of their "convectively activeness". The first three types are "convectively active": WS1 is the most vigorous deep convection associated with the larger mesoscale systems (large amounts of deep convective and cirrostratus clouds), WS2 is less vigorous (less deep convective and cirrostratus clouds) but exhibits more thick cirrus, WS3 exhibits somewhat lower cloud tops at medium optical depths but also somewhat more deep convective clouds with very high tops, suggesting isolated, smaller-scale convective systems with tops at a range of altitudes. The last three types are "convectively inactive': WS4 is primarily thin cirrus that is either outflow from distant convection or generated in isolation by other atmospheric motions, WS5 is a mixture of trade and shallow cumulus with some thin cirrus and WS6 is the marine stratus. They investigated where in the tropics each weather state is more likely to occur using 21.5 years of data, which is shown in Fig.45. The two most convectively active weather states (WS1 and WS2) occur the most frequently in the longitudes between 100°E and 180°E (warm pool), where WS3 is found predominantly over land areas. WS4 is concentrated in between the ITCZ and the SPCZ (South Pacific Convergence Zone) or north and south of the ITCZ. In the subtropics, the stratus type of WS6 is closer to the west coastal regions while the dominate type transforms to a more convective cumulus WS5 as it moves to the middle of the oceans.

3.3.2 Surface Radiative Fluxes

Using the analyzed ISCCP cloud properties as input variables, Zhang et al. (2004) have computed the radiative fluxes at TOA, in the atmosphere and at the surface. By comparing the difference of the fluxes between clear-sky and cloudy conditions, the effect of clouds (often called cloud forcing) can be evaluated. The shortwave, longwave and total net radiative fluxes at surface and the cloud effect on them are shown in Figs.46-48 using 3-hourly ISCCP data from 1983 to 2004. For shortwave (Fig.46), the full-sky net downward flux has its maximum in the tropics and very small values (near zero) near the poles. Cloud effect is to reduce the downward shortwave flux, and the greatest reduction occurs in the ITCZ, mid latitude storm track and west coastal stratus deck regions. For longwave (Fig.47), the full-sky net downward flux is negative (cooling the surface) globally, whereas its latitudinal variation is much small compared to that of shortwave. Clouds, on the other hand, warm the surface particularly in the higher latitudes. As a result, the total net radiation (Fig.48) shows a warming to the surface at the low latitudes while cooling at the high latitudes – this radiation-caused energy imbalance will then be neutralized by atmospheric and oceanic circulations. For the radiative effect of clouds on surface, they generally cause cooling for most of low and mid latitudes (±60°) and warming near the poles. ISCCP radiative fluxes data may be obtained from: http://isccp.giss.nasa.gov/projects/browse_fc.html.

3.3.3 Surface Turbulent Fluxes

As shown in Fig.37, the radiative energy imbalance at the surface is neutralized by net upward transport of sensible and latent heat by turbulence, the former arises from the temperature difference between surface and the air just above it, and the latter is due to evapotranspiration of water. A common method to estimate turbulent fluxes from an *in situ* platform is to measure the covariances of vertical velocity fluctuations with those of temperature and moisture (e.g., Fairall et al., 1997). From satellite, however, it is (at least as of today) impossible to determine these covariances, thus aerodynamic bulk formula (Fairall et al., 1996) has been used with satellite measured quantities (e.g., winds, SST, and near surface humidity etc.) as input variables. Based on bulk formula, the sensible and latent heat fluxes can be written as

$$\begin{aligned} F_{sh} &= \rho_a c_p C_H U(\theta_s - \theta_a), \\ F_{lh} &= \rho_a L_v C_E U(q_s - q_a) \end{aligned} \tag{26}$$

where ρ_a is the density of air; c_p is the specific heat of air at constant pressure; L_v is the latent heat of vaporization; C_H and C_E are the turbulent exchange coefficients for sensible and latent heat, respectively. The following variables may be derived from satellite measurements: U is the near-surface wind speed (commonly at 10 m above surface); θ is the potential temperature; q is the specific humidity where the subscripts s and a denote, respectively, surface and near-surface air. Under the auspices of the World Climate Research Programme (WCRP) Global Energy and Water Experiment (GEWEX) radiation panel, a project called SEAFLUX (Curry et al., 2004) has been established to evaluate and improve satellite-based surface turbulent fluxes over global oceans. Several published algorithms are currently being evaluated under the SEAFLUX project, including HOAPS (Hamburg Ocean Atmosphere Parameters and Fluxes from Satellite Data, Bentamy et al., 2003), J-OFURO (Japanese Ocean Flux Data Sets with Use of Remote-sensing Observations, Kubota et al., 2002), and GSSTF-2 (Goddard Satellite-Based Surface Turbulent Fluxes version 2, Chou et al., 2003). In general, these algorithms use satellite microwave observations to derive U and q_a (or q_s-q_a), and infrared observations to derive SST (or θ_s). For θ_a, generally there is no direct satellite measurement. It is usually derived from global analysis data, or from q_a by assuming a fixed value of relative humidity near surface, or from SST by assuming a fixed value of difference between surface and near-surface air temperature.

In the following, we show some examples of annual mean maps of the sensible and latent heat fluxes derived by HOAPS using data from 1988 to 2005 (figures from http://www.hoaps.zmaw.de). In Fig.49 are shown the fields of (a) SST (Kenneth, 2004), (b) 10-m wind speed and (c) 10-m air specific humidity (Bentamy et al., 2003) derived from satellite data from 1988 to 2005. The wind speed algorithm (unpublished yet) uses a neural network to derive the wind speed at 10 m height above the sea surface from SSM/I measurements. It consists of 3 layers: an input layer with 5 neurons (brightness temperatures of 19 GHz vertical, 19 GHz horizontal, 22 GHz vertical, 37 GHz vertical and 37 horizontal polarizations of SSM/I), a hidden layer with 3 neurons and an output layer with one neuron (wind speed). The network was trained with a composite dataset of buoy measurements and radiative transfer simulations. SST and near-surface specific humidity have a similar pattern – having greater values near the equator and decreasing toward to higher latitudes, with the maxima in the equatorial Pacific warm pool. Wind speed, on the other hand, has larger values in the higher latitudes, with the minimum occurring in the warm pool region. Surface sensible and latent heat fluxes are calculated using Eq.(26) and their multi-year mean maps are shown in Fig.50. In the tropics, sensible heat is about an order smaller than latent in magnitude, and negative values (heat transfer from atmosphere to ocean) cover most of tropical oceans. Regions where there is clear ocean-to-atmosphere sensible heat transfer can be found in the Gulf Stream, Kuroshio and southern hemisphere mid latitude regions. The largest values of latent heat fluxes are observed in the subtropical regions, with maxima as high as ~200 W m^{-2} for some regions. Note that latent heat fluxes are relatively low in the Pacific warm pool region due to low wind speed (Fig.49b) and high near-surface specific humidity (Fig. 49c).

4. Conclusion

In this article, we first reviewed the theoretical bases of remote sensing of various meteorological parameters from satellite. The review starts with solving radiative transfer equations under different conditions, followed by explaining the principles by which certain meteorological parameters can be retrieved. No attempts are made to describe a specific retrieval algorithm; the discussions on parameter retrieval were focused on principle rather than detailed procedures. Being the most relevant to tropical meteorology, the following retrieval problems are described. Under clear-sky conditions, surface temperature may be derived from window channels, and atmospheric temperature and moisture profiles may be retrieved from sounding channels in the infrared and/or microwave spectrum. Cloud properties, such optical depth and effective radius, can be retrieved from reflected solar radiation at visible and near-infrared channels. Observations in the microwave spectrum can infer many meteorological variables, such as precipitable water, cloud liquid, cloud ice water, rainfall, and ocean surface wind speed. Active sensors, such as cloud and precipitation radars, can provide the vertical distribution of hydrometeors. Scatterometers, on the other hand, can measure both ocean surface speed and direction.

Three topics are selected to demonstrate the applications of satellite observations in the tropics: hurricane studies, tropical rainfall and energy exchange at ocean surface. First, studies are introduced on the usage of satellite data to determine tropical cyclone's intensity, analyze its cloud and rainfall structure, and evaluate its water balance. For tropical rainfall, works mainly based TRMM satellite observations are reviewed, including rainfall climatology, horizontal and vertical structures of rain field, diurnal variation of rainfall and rainfall induced latent heating. Finally, researches on deriving radiative and turbulent fluxes at ocean surface are introduced. Because cloud is an important factor in influencing these fluxes, we first introduced satellite data derived cloud properties, with an emphasis on the efforts made by the ISCCP project.

Satellite data have been used in almost every subject in atmospheric studies – from studying the structure and evolution of a single storm to evaluating global climatology and long-term trend of geophysical variables, from pattern recognition of a phenomenon to quantitative retrieval, from using as numerical prediction model inputs to using as truth for model validation. It is practically impossible to all the topics in any one article or one book. Therefore, this article is in no way to be inclusive, but rather the author did "cheery-picking" from the vast volume of information – selecting topics that the author believes the most relevant to tropical meteorology. There is no doubt that some important works have been left unmentioned here. Interested readers are encouraged to follow the references listed in the Bibliography section for further information.

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Glossary

Absorption: A process in which incoming electromagnetic energy is transferred to the internal energy by gases, particles or surfaces. Gas absorption is caused by the changes of electron distribution, and vibrations and/or rotations of atoms and molecules. It occurs at distinctive wavelengths for a given gas molecule. The magnitude of imaginary portion of the dielectric constant for a liquid/solid material indicates its ability of absorption – a larger value means stronger absorbability.

Brightness temperature: The intensity emitted by an object as expressed in units of temperature. It is the temperature required to match the measured intensity to the Planck blackbody function at the given wavelength.

Radiometer: A device that passively measures the energy emitted from the target. The difference between a radiometer and a radar is that a radar transmits and then receives reflected energy, but a radiometer only receives energy naturally emitted by the target.

Radiative transfer equation: The equation that governs the flows of radiative energy. In the atmosphere, the radiative energy can be absorbed, emitted and/or scattered by gases, aerosols and hydrometeors. At the surface, radiative energy can be absorbed, emitted and/or reflected by land/ocean.

Remote sensing: Making a measurement without in touch with the target. Remote sensors can be passive (radiometers) or active (radars). The methods of remote sensing can be based on the emission, scattering and/or extinction of energy by the target.

Scattering: When incoming radiative energy interacts with a target, the target may divert part (or all) of the incoming energy into various directions without changing its frequency. This process is called scattering. When the target is very small in size compared to the incoming wave's wavelength, the scattering can be approximated by Rayleigh scattering. When the target is a sphere, the scattered electric field can be expressed by Mie theory. Generally, the larger the particle is, the more dominantly the scattered energy is distributed in the forward direction.

Sounding channels: Radiation measurements near (or at) the center of absorption lines can be used for deriving profiles of atmospheric temperature and/or water vapor. The channels used in satellite sensors near absorption lines are therefore called sounding channels.

Window channels: At some wavelength, the atmosphere is nearly transparent, i.e., no gas absorptions except for water vapor continuum (a small but broadly spread absorption by water vapor). AVHRR 11 and 12 μm channels are the most commonly used window channels for SST retrievals.

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