Chapter 5

SATELLITE OBSERVATIONS OF RADIATION AND CLOUDS TO DIAGNOSE ENERGY EXCHANGES IN THE CLIMATE: PART I

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5.1 Introduction: Energy cycles of the climate

The basic energy cycle that establishes the climate of our planet is more complicated than a simple balance between the solar radiation (wavelengths from $0.4 - 4\,\mu m$) absorbed by Earth and the terrestrial radiation (wavelengths from $4 - 100\,\mu m$) emitted when the temperature of Earth reaches equilibrium. The first complication is that the spherical shape of Earth leads to an equator-to-pole difference in solar heating; the resulting temperature contrast forces motions in the atmosphere and ocean that transport some of the energy from low to high latitudes. Thus, the radiation balance at the top of the atmosphere exhibits a net heating (absorbed solar flux exceeds emitted terrestrial flux) at low latitudes and a net cooling (emitted terrestrial flux exceeds absorbed solar flux) at high latitudes which is balanced by the dynamic transports. However, the global average radiation is essentially in balance when averaged over a whole year. Thus, the solar and terrestrial radiative fluxes are the drivers for all of the phenomena we call climate.

The second complication is that Earth's atmosphere is nearly transparent to solar radiation and nearly opaque to terrestrial radiation. Thus, most of the absorbed solar flux is deposited at the surface, but the terrestrial radiation emitted by the surface cannot escape easily to space. Consequently, the surface temperature must be larger to attain equilibrium - this is the "greenhouse effect". Another

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consequence of this fact is that the radiation balance at the surface is net heating, while that of the atmosphere is net cooling, setting up a vertical exchange of energy by other means.

The major cause of the opacity to terrestrial radiation is water vapor in the atmosphere, but water also creates the third complication in the energy cycle. Evaporation of water from the surface cools it; when some of this water vapor condenses to form precipitation, the atmosphere is heated. The actual process of cloud and precipitation formation also involves significant vertical re-distribution of energy as some water condenses at higher altitudes, falls and evaporates at lower altitudes and some water reaches the surface returning some energy to it. Since the water vapor can be transported by atmospheric motions, the surface cooling and atmospheric heating may not occur in the same places, implying an additional net transport of energy from warmer oceans to drier land and to colder latitudes. Thus, the energy balance at the surface is net radiative heating almost everywhere balanced by evaporative cooling and cooling by atmospheric and oceanic motions.

The water cycle also produces the fourth complication – clouds. Clouds change both the total amount of absorbed solar radiation and the total amount of emitted terrestrial radiation. In so doing, they also change the horizontal and vertical distributions of solar heating and terrestrial cooling, thereby altering the forcing for the atmospheric motions that transport heat and water. What makes this problem so challenging is that the clouds, too, are controlled by the atmospheric motions.

This basic understanding of the climate's energy cycle is accurate enough to describe the primary processes that control our environment, but not accurate enough to predict how the climate will respond to small perturbations, either natural or human-made. Predicting small transient responses or shifts of the climate equilibrium requires that we account for the changes in all of the processes that affect the radiation and water exchanges: at the center of all these exchange processes are clouds. The focus of this lecture is on the role of clouds in the radiation balance of Earth. Most of the results shown come from two major projects designed to develop the needed datasets: the International Satellite Cloud Climatology Project (ISCCP) and the Earth Radiation Budget Experiment (ERBE).

5.2 Remote sensing of clouds and radiation

Improved modeling of cloud effects on radiation is the desired outcome of the analysis of observations of clouds and radiation; however, the only practical way to obtain comprehensive measurements of cloud properties is to infer them from the radiation measurements, themselves. That is, the essence of the analysis of remote sensing data, whether from satellites, aircraft or the surface, is to model the detailed effects of clouds on radiation to link physical variables to the observed radiation (Rossow, 1989). Thus, development and improvement of radiative models including cloud effects must proceed in parallel with collection and analysis of measurements. This lecture will concern satellite observations, but the approach can be used for other types of data with some modifications.

Determination of specific cloud properties is best done with radiation measured at specific wavelengths that are selected to be more sensitive to the desired quantity and less sensitive to other quantities. However, since satellites are expensive to build and have been designed to measure other

Satellite Name	Instrument Name	Time Period	Area of Coverage	Comments
NIMBUS-4	IRIS	1970	Globe	The only IR spectra that are available
NIMBUS-7	TOMS	1978-1991	Globe	UV imager used to measure cloud amount and reflectivity
NIMBUS-7	THIR	1978-1985	Globe	IR imager used to measure cloud amount and height
NIMBUS-7	SMMR	1978-1985	Globe	Microwave scanner used to measure water vapor, cloud liquid water amount and precipitation rate
NOAA-6–12	AVHRR	1979-1991	Globe	Solar/IR imager used to measure cloud amount, optical depth, particle size, and height
	HIRS-2	1979-1991	Globe	IR sounder used to measure atmospheric temperature and water vapor profiles and cloud amount, height and optical depth
DMSP	SSM/I	1987-1991	Globe	Microwave scanner used to measure water vapor, cloud liquid water amount and precipitation rate
SMS-1,2	VISSR	1974-1982	Regional	Solar/IR imager can be used to measure cloud amount, optical depth and height
GOES-1–7	VISSR	1981-1991	Lower latitudes Americas	Solar/IR imager used to measure cloud amount, optical depth and height
	VAS		Lower latitudes Americas	IR sounder used to measure atmospheric temperature and water vapor profiles and cloud amount, height and optical depth
GMS-1–4	VISSR	1983-1991	Lower latitudes Western Pacific	Solar/IR imager used to measure cloud amount, optical depth and height
METEOSAT-1–5	VISSR	1979-1991	Lower latitudes Africa/Europe	Solar/IR imager used to measure cloud amount, optical depth and height

Table 5.1: *Some satellite datasets that can be or are being used to study clouds.*

aspects of the atmosphere and surface as well, the best choices of wavelengths are not always available. Currently available datasets that provide some information about cloud properties and their variations are listed in Table 5.1 by satellite and instrument name. Also given are the time periods covered and a simple indication of the portion of the globe covered. This list is not historically complete, rather it mentions datasets that are in good condition and readily available.

Although many cloud properties affect the radiation at any one wavelength, these properties can be arranged in order of decreasing magnitude of their effect. With a limited number of wavelengths measured, we can only measure a few of the most important properties by specifying the remaining quantities. The extent to which the specified properties vary determines the uncertainty in the measured properties.

The four most important cloud properties that can be measured from available satellite datasets are top temperature, optical thickness, particle size and liquid water content (I will discuss cloud "amount" later). Cloud top temperature can be thought of as the actual physical temperature of the topmost part of a cloud; when the cloud is opaque to radiation, then the satellite "sees" an emitted radiation equivalent to a black body at this temperature. However, when the cloud is not opaque, the radiation emitted by the cloud may originate within it and may be added to by radiation passing through the cloud from below. In other words, an optically thin cloud would appear to emit radiation equivalent to a temperature higher than that of its top (since lower atmospheric levels and the surface are usually warmer than clouds). Measurements at infrared wavelengths (usually near $10 - 12\,\mu m$) are most commonly used to measure cloud top temperature (e.g., Minnis and Harrison, 1984; Stowe et al., 1988; Rossow et al., 1989 and references therein); however, some microwave wavelengths measure the temperature at the top of the layer containing ice or rainfall (Mugnai and Smith, 1988).

During daytime, the optical thickness of the cloud can be measured by the amount of sunlight reflected by the cloud (e.g., Rossow et al., 1989). This method is not very sensitive to lower values, but covers the whole range of values that occur. Day or night, lower values of cloud optical thickness can be estimated from the shape of the spectrum of infrared radiation emitted by the combination of cloud, atmosphere and surface, but this method does not work for larger values. (Infrared spectrometer measurements of Earth have been obtained only in 1970 even though such measurements are "routine" for spacecraft missions to other planets.) When the optical thickness of the cloud can be measured, then its effect on the measurement of cloud top temperature can also be removed (Rossow et al., 1989; Minnis et al., 1990); however, for ice clouds the relation between the optical thickness at solar and infrared wavelengths is more uncertain because it depends on particle shape (Wielicki et al., 1990). Once cloud top temperature is known, the location of cloud top in the atmosphere can be determined by comparing this value to the profile of atmospheric temperature. IR sounder measurements, on the other hand, can be used to infer the cloud top pressure more directly; when the temperature at that pressure differs from the apparent radiometric temperature, the infrared optical thickness can also be inferred (Wylie and Menzel 1989).

The average distribution of cloud top pressures for January and July is illustrated in Fig. 5.1 using results from ISCCP; Fig. 5.2 shows the annual mean latitudinal distribution of cloud optical thicknesses and top temperature from ISCCP. The cloudiest latitudes constitute the tropical and

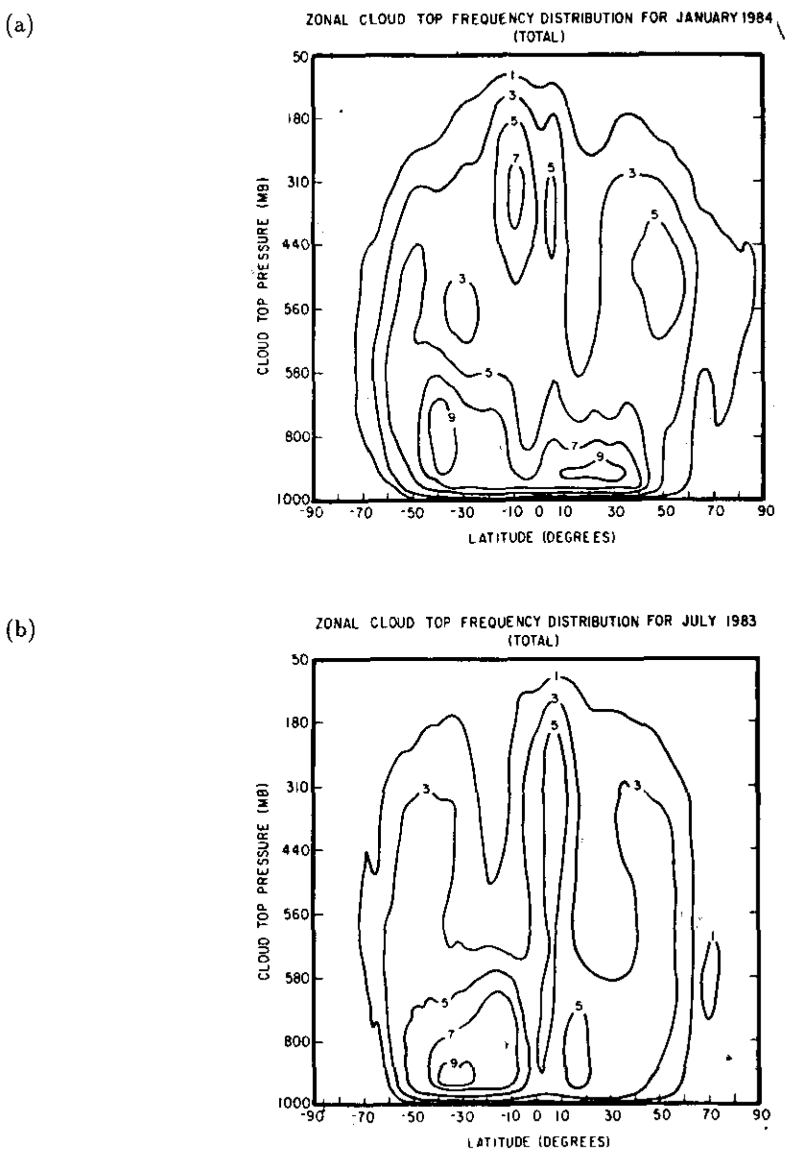

Figure 5.1: *(a) Zonal mean frequency of cloud top pressure for January 1984 from ISCCP, (b) Zonal mean frequency of cloud top pressure for July 1983 from ISCCP.*

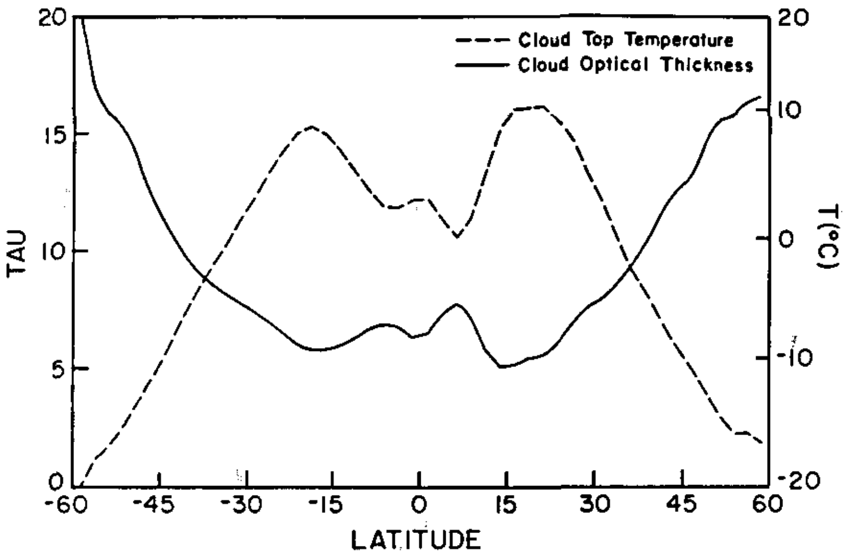

Figure 5.2: *Zonal, annual mean cloud optical thickness and top temperature* (K) *from ISCCP.*

midlatitude storm zones. The highest clouds occur in the summer hemisphere part of the tropics, whereas there are more high clouds in the winter midlatitudes (Fig. 5.1). Despite larger cloud top pressures in midlatitudes than in the tropics, cloud top temperatures are much colder at higher latitudes because the atmospheric temperatures are lower at a particular pressure level (Fig. 5.2). The northern hemisphere exhibits larger seasonal variations than the southern hemisphere. Cloud optical thicknesses are larger in the summer tropics and winter midlatitudes than in the opposite seasons; the largest optical thicknesses occur at higher latitudes (Fig. 5.2).

Cloud particle size affects both the amount of sunlight reflected by a cloud and the shape of the spectrum of infrared radiation it emits. Thus, additional measurements at these wavelengths can be used to infer the cloud particle size, together with the optical thickness and top temperature (e.g., Arking and Childs, 1985; Nakajima and King, 1990; Ackerman et al., 1990; Wielicki et al., 1990; Wetzel and Vonder Haar, 1991). Some preliminary results for low level, liquid water clouds are illustrated in Fig. 5.3, showing a systematic difference in average particle size between clouds over land and oceans: average particle radius is about 12 μm over oceans and about 10 μm over land.

Measurements of cloud optical thickness and particle size can be used to estimate the amount of water in a cloud. Microwave radiation at certain wavelengths can be used to measure more directly the total amount of liquid water over oceans by measuring the amount of radiation that is absorbed by the cloud (eg., Njoku, 1982; Alishouse et al., 1990; Curry et al., 1990). Figure 5.4 shows a comparison of these two measurements; the discrepancy in these results depends on the fact that the two measurements have different sensitivities to lower values. If the optical thickness values from visible wavelength measurements are truncated at about 3, that is, if the thinner clouds are removed from the ISCCP results, then the agreement between the two results is as shown; without this truncation the ISCCP results are about a factor of 2 – 5 lower.

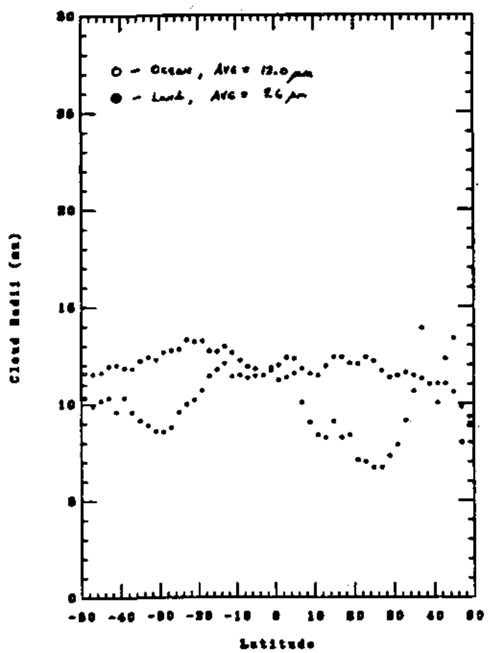

Figure 5.3: *Zonal mean cloud particle radius* (μm) *for warm clouds (thesis work of Q.-Y. Han).*

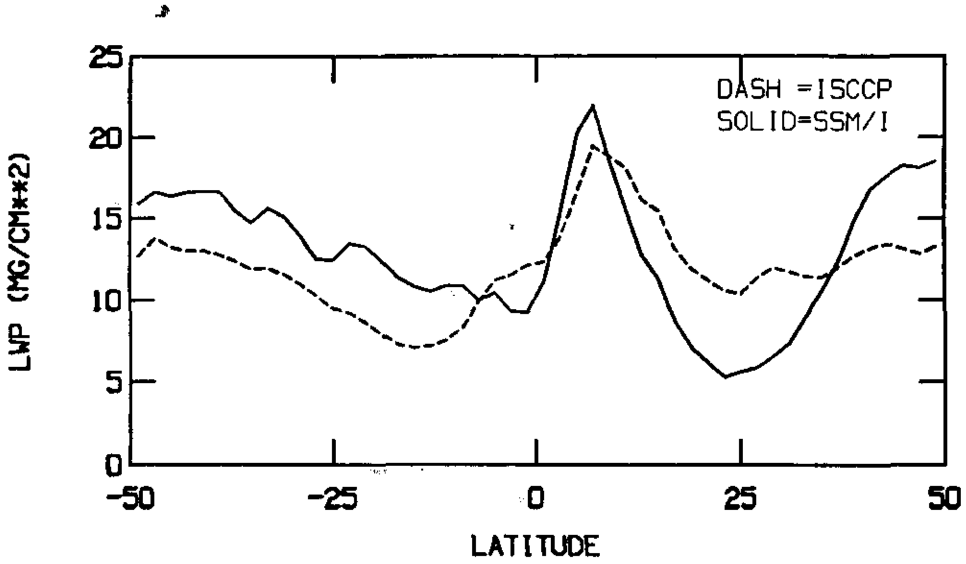

Figure 5.4: *Comparison of zonal, monthly mean liquid water path* ($mg\,cm^{-2}$) *inferred from SSM/I measurements and ISCCP optical thickness values (from thesis work of B. Lin).*

The accuracy with which these cloud properties can be measured depends both on the fidelity of the radiative transfer model used in the analysis and on the amount of information provided in the satellite dataset. Measurement of radiation at only one or two wavelengths limits the analysis to only one or two quantities; the dependence of the measurements on other quantities must be assumed. Measurement of more wavelengths eliminates these assumptions and improves the accuracy of the results. Future multi-wavelength datasets should produce more accurate results; however, as will be illustrated, these results already have enough accuracy to begin study of the key climate problems.

There are also some problems in performing the analyses described above that need more study. The most significant problem concerns how to account for the effects of horizontal inhomogeneities in clouds at "smaller" scales. The question that needs answering is whether there are scales at which the radiative effects of the cloud can be represented by spatially averaged properties and the effects of spatial variations at smaller scales neglected or treated in some "statistical" fashion. There are three considerations:

1. cloud properties vary on scales much smaller than those on which the radiation interacts with the cloud,
2. cloud properties exhibit a spectrum of variation that mirrors that of the atmospheric motions, with larger amplitudes at larger scales, and
3. for each spatial scale there is a characteristic time scale at which the variations occur.

The first consideration means that cloud variations at really small scales do not actually alter the radiation and the second consideration suggests that the magnitude of the cloud variations, and therefore, their importance to the radiation, is much smaller at these smaller scales. The third consideration enters when one considers that the radiative response time of the atmosphere and ocean is much larger than the time for changes of clouds at smaller scales; hence, the importance of the smaller spatial scale variations may be lessened.

This problem is usually cast in terms of determining the cloud "amount", which is thought of as the fraction of the area covered by cloud. The difficulty is that, while this concept seems straightforward, it is actually difficult in practice to define what is meant by "cloud" and where, precisely, the boundary between "cloud" and "non-cloud" is (Wielicki and Parker, 1992). Earth's atmosphere contains some particles everywhere at all times, so that a division of the atmosphere into these two parts requires some "threshold" optical thickness, as in: if the particle optical thickness is larger than 0.1, then it is cloudy. Cloud "cover" defined this way is essentially a radiative quantity that depends directly on the value of this threshold and remains, in a sense, arbitrary. The presence of a spectrum of scales of cloud variability has led to studies representing cloud cover as a fractal (e.g., Lovejoy and Schertzer 1990). Although this mathematical technique provides a valuable new way to represent this complex spectrum of variations, it does not necessarily solve any problems when used in this form. Instead, it may be better to think of "non-cloud" simply as "cloud" with small or zero optical thickness. In this form, the problem becomes the problem of treating the effects of the smaller scales of variability on the larger scale radiation fluxes. The above arguments imply that if the threshold used to divide "cloud" from "non-cloud" is well defined and small enough, then the accuracy of radiative effects

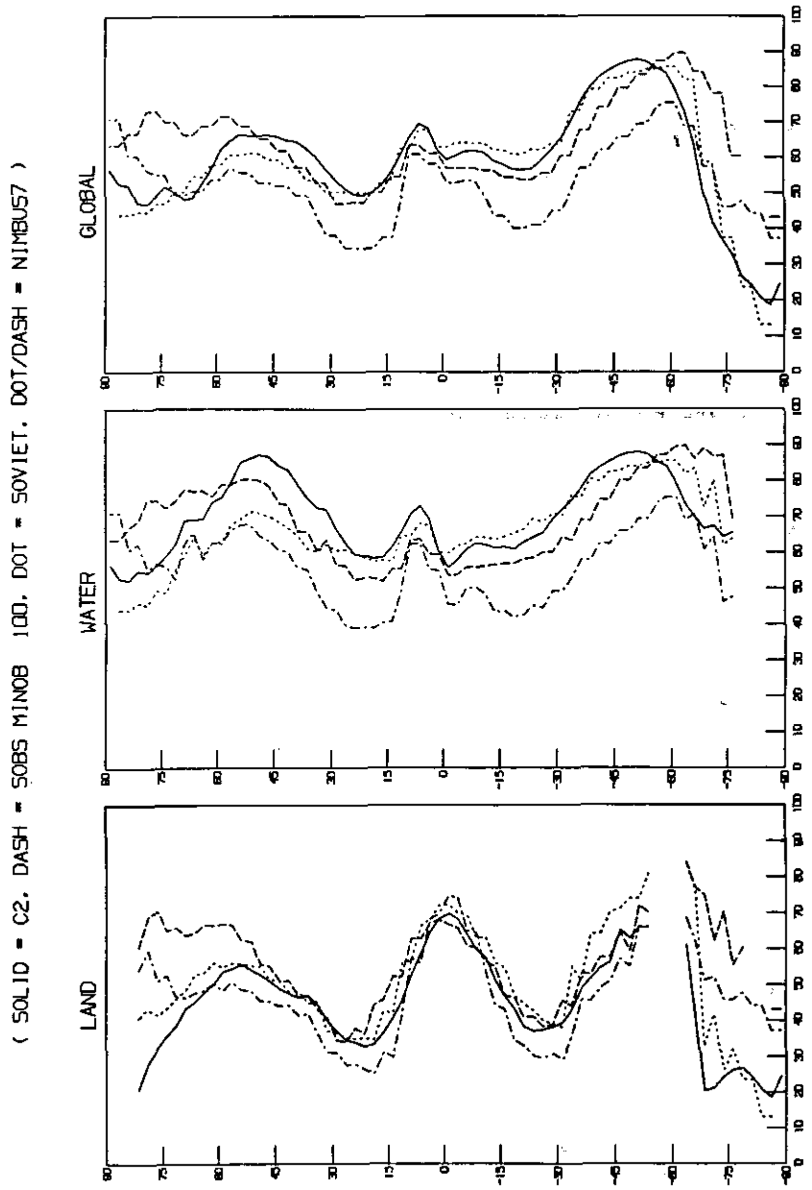

Figure 5.5: *Comparisons of zonal, annual mean cloud amounts (%) from ISCCP, surface observations, METEOR satellite and NIMBUS-7.*

calculated with this value will be high. However, such a "cloud amount" is most meaningful when the corresponding cloud optical properties are reported.

Figure 5.5 compares the average cloud amounts inferred from three different satellite systems (Rossow and Schiffer, 1991; Stowe et al., 1989, Mokhov, 1992) with those observed by surface weather observers (Warren et al., 1986, 1988). The good correspondence of these results, based on different definitions of the detection threshold, encourages one to accept that this quantity is being measured with useful accuracy. In particular, the agreement between ISCCP and surface observers is quite good and, when the ISCCP cloud detection sensitivity is reduced, cloud amounts similar to the METEOR and NIMBUS-7 results are obtained. The satellite datasets have better coverage and sampling over oceans than the surface dataset.

A more detailed comparison between individual ISCCP results and surface observations reveals that the usual distribution of differences (Fig. 5.6(a)) is actually more complex when displayed like Fig. 5.6(b). The more complex pattern is actually to be expected when comparing a "large-area-like" result like that of ISCCP with the "point-like" observation of the ground observer. When the former shows either completely clear or completely overcast conditions, which occurs about half the time, the latter shows good agreement. The disagreement at zero cloud amount and maximum difference (lower left corner of Fig. 5.6(b)) is associated with a small number of clouds that are missed by the satellite detection. When the satellite dataset indicates partial coverage, there are two different kinds of situations: one where a relatively uniform scattering of small clouds covers the whole area and one where some portion of the area is overcast. The first case explains the distribution of values near zero difference for all cloud amounts in Fig. 5.6(b), indicating good agreement for scattered cloudiness. The second case explains the values representing the maximum disagreement in Fig. 5.6(b):

1. when the fraction of the area covered by clouds is small, the probability is high that a single surface station will be under clear conditions, so that the cloud amount difference is in the upper branch and

2. when the fraction of the area that is overcast is large, the probability is high that a single surface station will be under overcast conditions, so that the cloud amount difference is in the lower branch.

At 50% coverage, both branches have equal probability. The results in Fig. 5.6 exhibit this behavior with a small shift that indicates that the ISCCP results underestimate cloud amounts over land by about 5%.

Another specific and difficult problem in the design of good radiative transfer models for radiation in cloudy atmospheres is how to treat the effects of non-spherical cloud particle shapes. The most sophisticated theories and models for radiation in a cloud are based on the assumption of spherical particle shape (Hansen and Travis 1974), which is an excellent assumption for liquid water clouds, is less accurate for rainfall, and is incorrect for most ice clouds. The shape of the cloud particles affects the detailed angular distribution of radiation that they scatter (e.g., Asano and Sato, 1980; Takano and Liou, 1989). In optically thicker clouds, where each photon is scattered many times, the shape effects are less important; however, we currently lack a general theory for treating this problem.

5.3 Calculating radiative fluxes

The most direct way to measure the total radiative fluxes (summed over all wavelengths) is with instruments that measure radiation at all wavelengths: one type makes a single measurement of

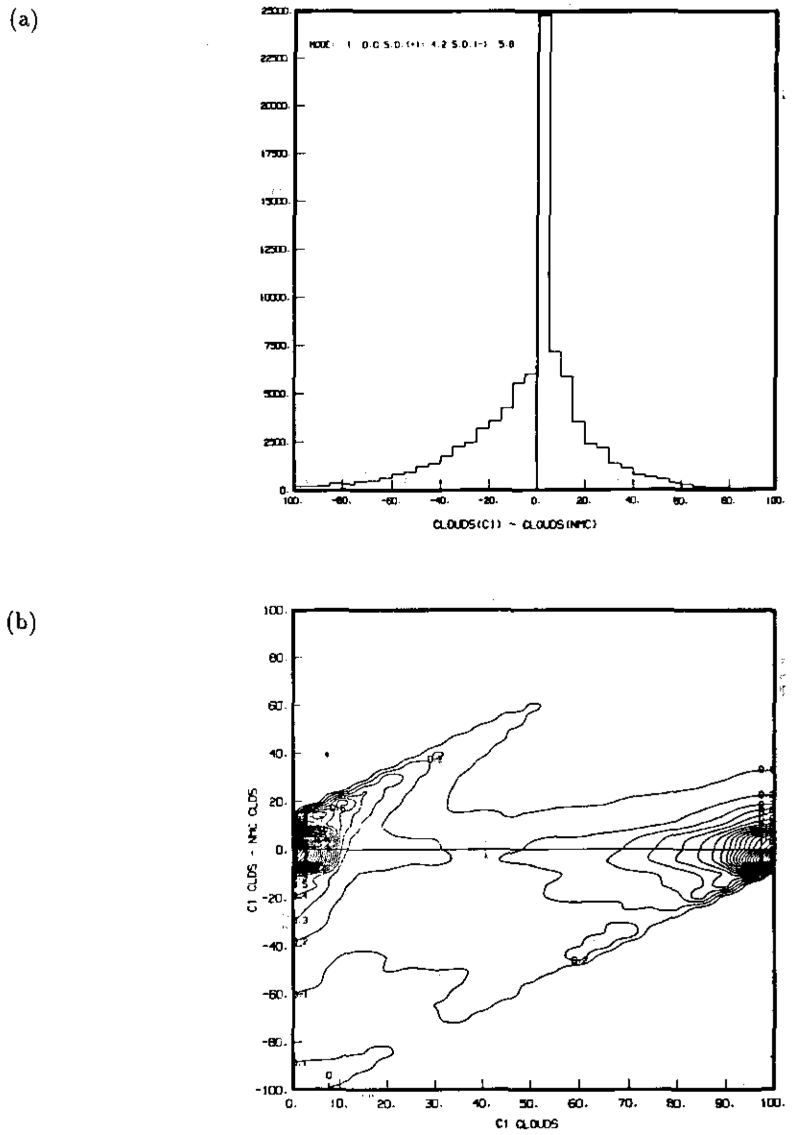

Figure 5.6: *(a) Frequency distribution of the differences between ISCCP and surface observed cloud amounts from the global distribution of weather stations for July 1985, (b) Same data shown in the upper figure as a function of ISCCP cloud amount (%). Contours indicate frequency as a fraction of the peak value.*

the radiation integrated over all wavelengths (called a "broadband" measurement) and another type makes measurements at many discrete wavelengths that can be added together to get the total. The former type of measurement has been favored (Jacobowitz et al., 1984; Barkstrom and Smith, 1986), generally because it was somewhat more difficult to calibrate instruments that make multi-wavelength measurements; however, certain types of spectrometers that make measurements that are continuous in wavelength, such as the IRIS for infrared wavelengths, may actually be easier to calibrate; but almost no spectroscopic observations of Earth have been made from space. Estimates of total fluxes have also been made based on measurements of a smaller number of wavelengths within the spectral range (Raschke et al., 1973; Gruber, 1977). Table 5.2 summarizes the datasets that have (or could be) used for this purpose.

Satellite Name	Instrument Name	Time Period	Area of Coverage	Comments
NIMBUS-7	Scanner	1978-1981	Globe	Broadband
	MFOV	1978-1990	Globe	Broadband
	WFOV	1978-1990	Globe	Broadband
NOAA-6–12	AVHRR	1979-1991	Globe	Narrowband measurements used to estimate radiation for total spectrum
	HIRS-2	1979-1991	Globe	Multi-wavelength IR measurements used to estimate radiation for total IR spectrum
ERBS (ERBE)	Scanner	1984-1990	Lat < 60°	Broadband
	WFOV	1984-1991	Lat < 60°	Broadband
NOAA-9 (ERBE)	Scanner	1985-1987	Globe	Broadband
	WFOV	1985-1991	Globe	Broadband
NOAA-10 (ERBE)	Scanner	1986-1989	Globe	Broadband
	WFOV	1986-1991	Globe	Broadband

Table 5.2: *Some satellite datasets that can be or are being used to study radiative fluxes at the top of the atmosphere and at the surface of Earth.*

In all cases, these satellite measurements are *radiances* not *radiative fluxes*, the latter is obtained by integrating the former over all angles. Since our atmosphere is a very thin shell around the planet, we usually divide the radiation going in all directions into "upward" and "downward" parts, where each part represents an integration of the radiation over a hemisphere. Since the atmosphere and surface are not homogeneous in their properties and since scattering processes, in particular, do not distribute radiation uniformly in all directions, the radiation intensity necessarily varies with the direction in which the satellite sensor points. The nature of this variation also varies strongly with wavelength. The vertical inhomogeneity of the atmosphere naturally divides the problem of inferring fluxes into two horizontal scales: one much larger than the depth of the atmosphere and one smaller than or similar to the depth of the atmosphere. At smaller scales the problem involves a three-

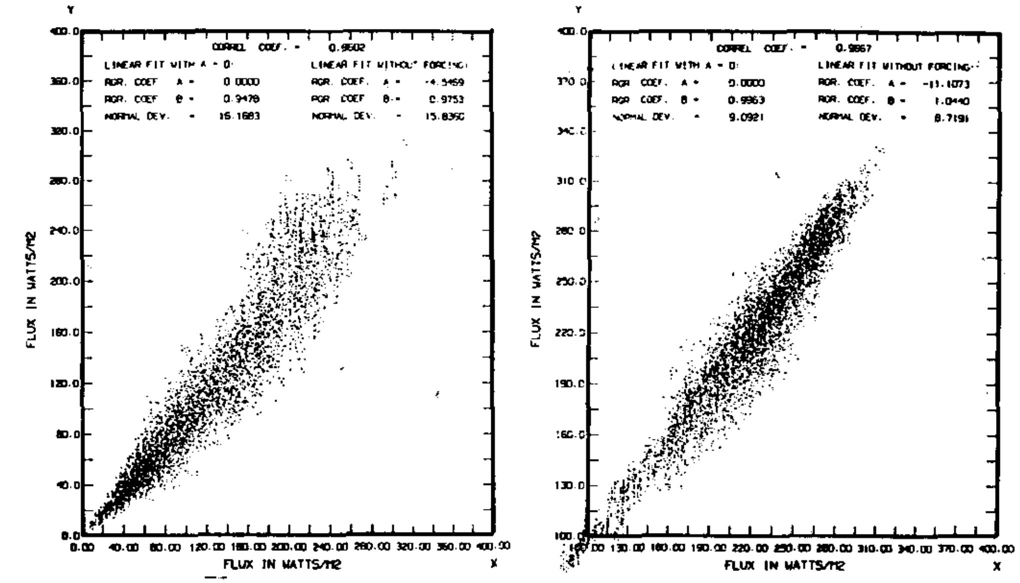

Figure 5.7: *Scatter-plot of daily values of ERBE against ISCCP against solar (left) and terrestrial (right) fluxes over the globe.*

dimensional radiation field, whereas the problem is two-dimensional at larger scales. For example, determining the "upward" flux for a region smaller than the vertical extent of the atmosphere is much more difficult because contributions from adjacent regions are just as important. Improving the treatment of the angular distribution of radiation to infer fluxes from radiance measurements is still a key problem.

Accurate determinations of total solar fluxes are made particularly difficult by the rapid daily variation of solar illumination: the fluxes change by about 1% to > 10% in about 30 minutes. Most satellite datasets provide very sparse samples of this diurnal variation; in fact, only the ERBE dataset contains measurements from more than one satellite, which still only gives a statistical sample over the diurnal cycle in one month (Brooks et al., 1986).

It is also possible to calculate the radiative fluxes, if all of the necessary physical variables are measured and a radiative model of sufficient accuracy is used (Rossow and Lacis, 1990). The variables that are needed are: surface spectral albedo and emissivity, surface temperature, profiles of atmospheric temperature, humidity and ozone, abundances of other radiatively important atmospheric gases that are uniformly mixed, aerosol optical thickness and optical constants, cloud optical thickness and its vertical distribution, as well as cloud particle size, shape and phase. These variables also must be available with sufficient space and time sampling to capture the important variations of the radiation fluxes. Attempts to infer the radiation fluxes in this way are being made using datasets from the International Satellite Cloud Climatology Project (ISCCP) and from operational weather observations. Although this approach requires much more data, including special datasets needed to validate the radiative model, it allows for an examination of the detailed effects of clouds and other

Quantity	ERBE[1]	NIMBUS-7[2]	ISCCP
Downward Solar	341.3	343.0	341.4
Reflected Solar	102.1	111.1 (102.5)	113.4
Cloud effect	-48.9	-50.9	-52.2
Emitted Terrestrial	234.5	233.7 (234.9)	232.4
Cloud effect	31.0	24.1	22.2
Net Flux	4.7	-1.9 (5.6)	-4.4
Cloud effect	-17.9	-26.8	-30.0

Table 5.3: *Comparison of annual mean radiative fluxes (in Wm^{-2}) at the top of the atmosphere measured by ERBE and NIMBUS-7 with those calculated using datasets from ISCCP. The "annual" mean values for ERBE and ISCCP are the average over April, July and October 1985 and January 1986; for NIMBUS-7 these are averages over summer (June, July and August) 1979 and winter (December 1979, January and February) 1980. The cloud effects are represented by differences between the total fluxes including clouds and clear fluxes excluding clouds. The net flux is the downward solar flux minus the reflected solar minus the emitted terrestrial flux.*

parts of the climate system on the radiation balance. Tables 5.3 and 5.4 and Fig. 5.7 compare the results calculated from ISCCP data with coincident ERBE measurements.

The calculated fluxes from ISCCP datasets agree with the direct measurements about as well as they agree with each other; Table 5.4 shows the statistics from scatterplots, like those shown in Fig. 5.7, for both daily mean and monthly mean values.

	Correlation	Regression Slope (ERBE/ISCCP)	Intercept [Wm^{-2}]	RMS Difference [Wm^{-2}]
DAILY STATISTICS				
Reflected Solar	0.96	0.95	0.0	16.6
Reflected Solar - Clr	0.95	0.99	0.0	14.7
Emitted Terrestrial	0.96	1.04	-7.5	9.0
Emitted Terrestrial - Clr	0.99	1.08	-12.1	5.5
MONTHLY STATISTICS				
Reflected Solar	0.99	0.95	0.0	9.2
Reflected Solar - Clr	0.94	0.93	0.0	14.3
Emitted Terrestrial	0.99	1.07	-15.6	4.0
Emitted Terrestrial - Clr	0.99	1.11	-20.1	3.7

Table 5.4: *Comparison of daily and monthly mean fluxes measured by ERBE with values calculated daily and monthly from ISCCP results for July 1985.*

[1] Values taken from ERBE data tapes.
[2] Values taken from Ardanuy et al. (1991); values in parentheses are from Kyle et al. (1990) but for the same months as ERBE data.

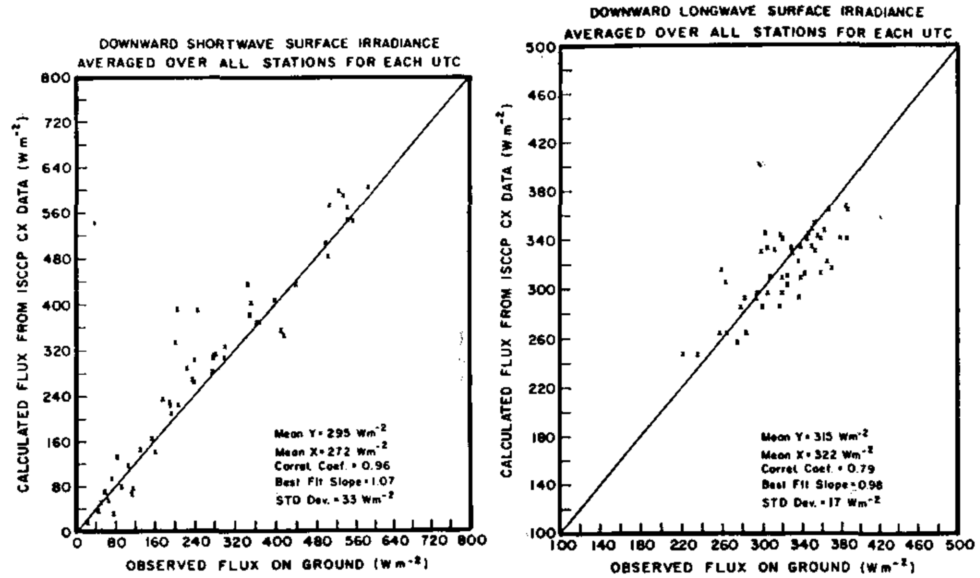

Figure 5.8: *Scatter-plot of ISCCP against surface measured downwelling solar (left) and terrestrial (right) fluxes for the Wisconsin SRB experiment.*

There is not as much data available to check the calculations of the fluxes at the surface using the same ISCCP datasets and radiative transfer model; however, Fig. 5.8 shows a comparison based on an intensive surface observation campaign conducted in support of an International Surface Radiation Budget project. The agreement is better for daily mean values:

1. for shortwave, the regression slope is 1.01 with an RMS difference of $10\,W\,m^{-2}$ and a bias of $4\,W\,m^{-2}$ and
2. for longwave, the regression slope is 0.98 with an RMS difference of $17\,W\,m^{-2}$ and a bias of $6\,W\,m^{-2}$.

5.4 Summary

Two kinds of radiation measurements have not been exploited for Earth observations as yet. One kind is measurement of continuous spectra over large portions of the electromagnetic spectrum. Although such measurements of other planets at infrared wavelengths have become common, only one spectrometer has ever been flown on an Earth- observing satellite, IRIS on NIMBUS-3 and 4 covering part of the terrestrial radiation spectrum from about $6-25\mu m$. Although such measurements are usually used to determine composition, they contain a rich amount of information about clouds, including particle properties and vertical structure (e.g., Carlson et al., 1992). Of particular interest for cloud studies would be infrared and microwave spectral measurements; infrared spectra could also be directly integrated to get the total terrestrial radiation fluxes.

The second kind of un-exploited measurement is of the polarization of reflected sunlight, which has been used to study the atmospheres of other planets (e.g., Hansen and Hovenier, 1974) but not Earth (but see Hansen, 1971). Polarization of microwave radiation is being used to examine the structure of precipitating cloud systems, but more theoretical work is required (Mugnai and Smith, 1988). The polarization of scattered solar radiation is particularly sensitive to the micro-scale properties of the scattering medium; hence it can be used to determine the cloud particle size distribution, shape and phase. It can also be used to obtain another estimate of cloud top pressure. Moreover, this type of measurement is much more sensitive to scattering at low optical thicknesses than measurements of the intensity of the reflected radiation, so that the determination of the properties of optically thin aerosols in Earth's atmosphere would be more accurate. Current plans for new instruments may produce these types of datasets within the next 5-10 years.

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