Chapter 6

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

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6.1 Cloud properties

Since the early part of this century, we have had a fairly accurate idea of the variations of cloud cover among different climate zones, except for the more remote portions of the tropical and southern oceans and polar regions (Hughes 1984). Standardization of cloud classification for surface weather observers after the Second World War has also provided a long record of changing cloud morphological types (Warren et al. 1986, 1988). However, quantitative information on the physical properties of clouds has been lacking, particularly for those properties that determine their influence on radiative energy exchanges. Recent analyses of satellite datasets , particularly by the International Satellite Cloud Climatology Project (ISCCP), are providing the first global surveys of some of these cloud properties (Rossow and Schiffer 1991). The results presented below come from ISCCP.

The average cloudiness of Earth is over 60%. Although earlier estimates gave a value slightly larger than 50%, the higher value is obtained when a proper sample of the southern oceans is included and when thinner cirrus clouds are included. (The precise value of "cloud" cover depends directly on how small the optical thickness cut-off value is: a value of 60 – 65% is associated with a cut-off of about

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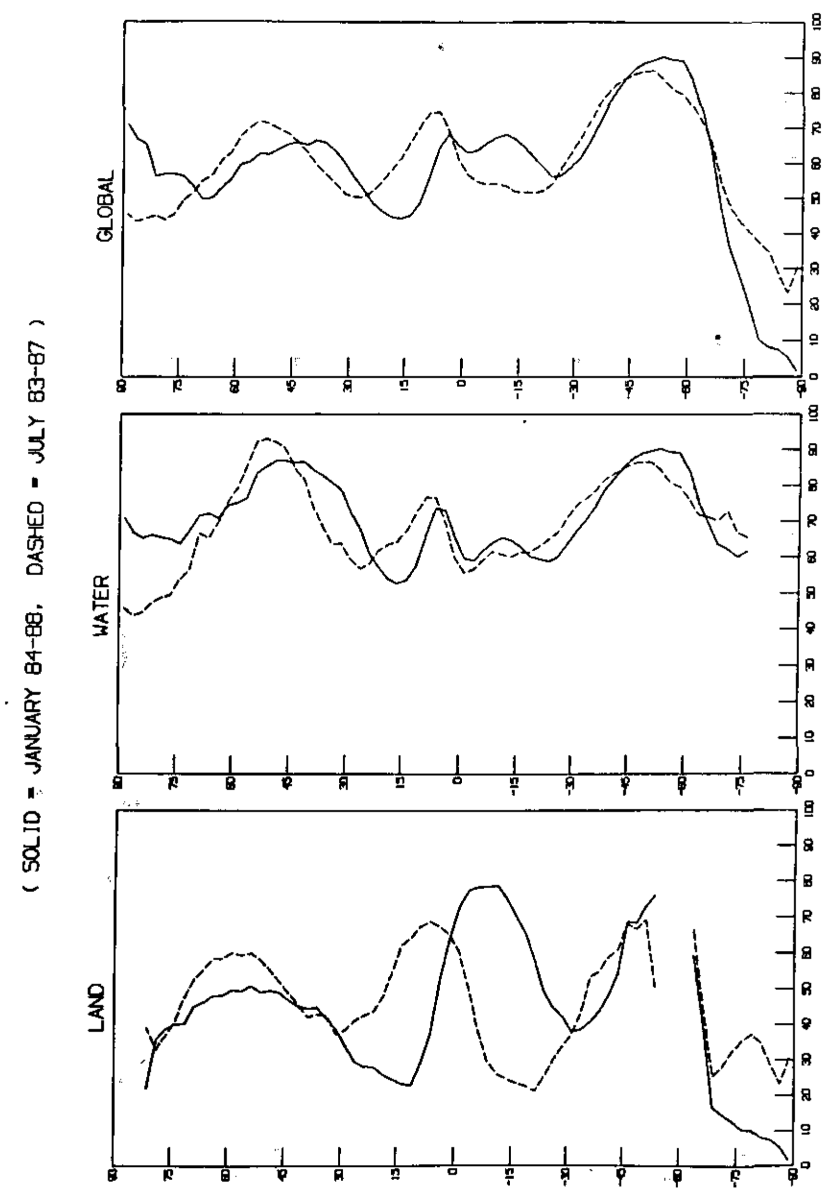


Figure 6.1: *Zonal, annual mean cloud amounts (%) from five years of ISCCP data for January and July.*

0.2. Clouds with smaller optical thicknesses would only alter radiative fluxes by a few percent. If the cut-off were reduced to about 0.1 or slightly less, global cloud cover would be nearly 100% at all times.) Figure 6.1 shows that the latitude zones containing stronger atmospheric motions ("storms") are generally cloudier (60 – 80%) than the more quiescent zones (40 – 50%). The larger scale storm systems in middle latitudes produce somewhat more extensive cloud cover than the smaller scale tropical systems. Drier land areas are generally less cloudy than the oceans at all latitudes, except in the Intertropical Convergence Zone (ITCZ). There are also large ocean "deserts" with little cloud cover. The sea-ice covered polar oceans are very cloudy, but the high altitude ice sheets are more like deserts. Variations in cloud cover with longitude are just as large as they are with latitude (Fig. 6.2).

The distributions of cloud optical thicknesses and cloud top pressures exhibit large variations that also coincide with the classical climate regimes (Fig. 5.1 and 5.2 in Part I). The "storm" zones have optically thicker clouds and more cloud tops at lower pressures (higher altitudes), while the quiescent zones have thinner clouds and fewer clouds in the upper troposphere. Clouds over oceans tend to be optically thinner and to occur more frequently at lower levels than over land (the latter effect is partly due to the higher topography of land areas and to a thicker and drier planetary boundary layer). This difference is also associated with a difference in the average cloud particle sizes for low-level clouds: those over oceans tend to have larger particles than over land (Wallace and Hobbs 1977) and to form precipitation more easily, which may explain their lower optical thicknesses (Tselioudis et al. 1992).

We can consider combinations of cloud radiative properties to represent different cloud types and examine how the relative frequency of occurrence of these cloud types varies with climate regime. One classification can be defined by cloud top pressure and optical thickness (Fig. 6.3(a)); these two properties, together with the frequency of occurrence, determine the primary effect of clouds on the solar and terrestrial radiative fluxes. The quiescent climate regimes over oceans are dominated by a single cloud type that is low-level and optically thin (Fig. 6.3(b)); over land, some thin, low-level clouds occur, but most clouds are thin, high-level cirrus. Midlatitudes have two common cloud types (Fig. 6.3(c)): one with middle to upper level cloud tops and large optical thicknesses, corresponding to storms, and one low-level, optically thin type corresponding to fair weather. Tropical latitudes have three cloud types (Fig. 6.3(d)): low-level, optically thin clouds associated with fair weather and two types associated with convective complexes. The convective "towers" have very high optical thicknesses and very high tops; the mesoscale anvils have moderate optical thicknesses and somewhat lower tops. Cirrus clouds are at similar altitudes to the convective clouds in the tropics but are optically thin. The latter cloud type can also occur alone in both the tropics and midlatitudes.

6.2 Seasonal variations

Low and high latitudes exhibit completely different seasonal variations in cloud properties. At low latitudes seasonal temperature variations are small and the seasons are marked by large changes in precipitation. The "summer" season is the wet season with large storms; the cloud cover becomes much more extensive and the cloud types associated with convective complexes occur much more

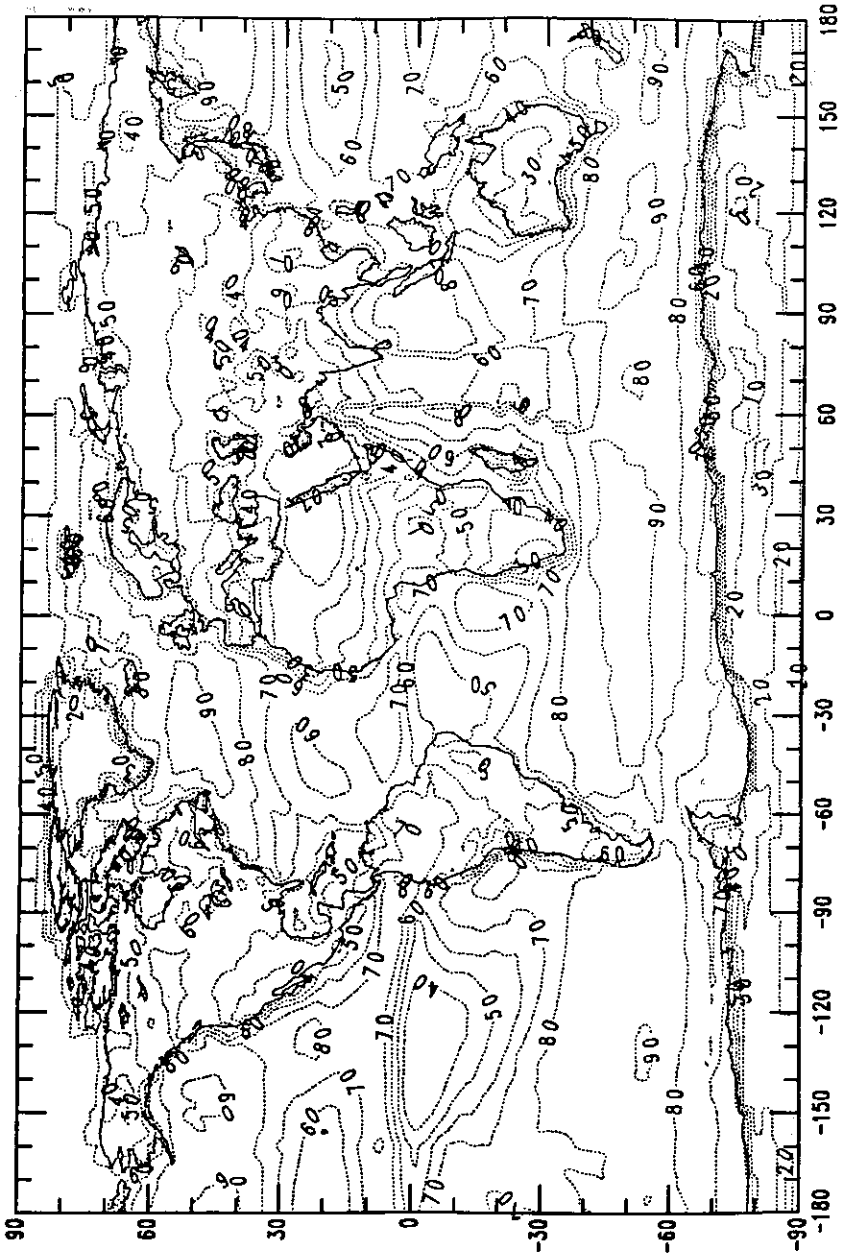


Figure 6.2: *Annual mean cloud amount (%) from five years of ISCCP results.*

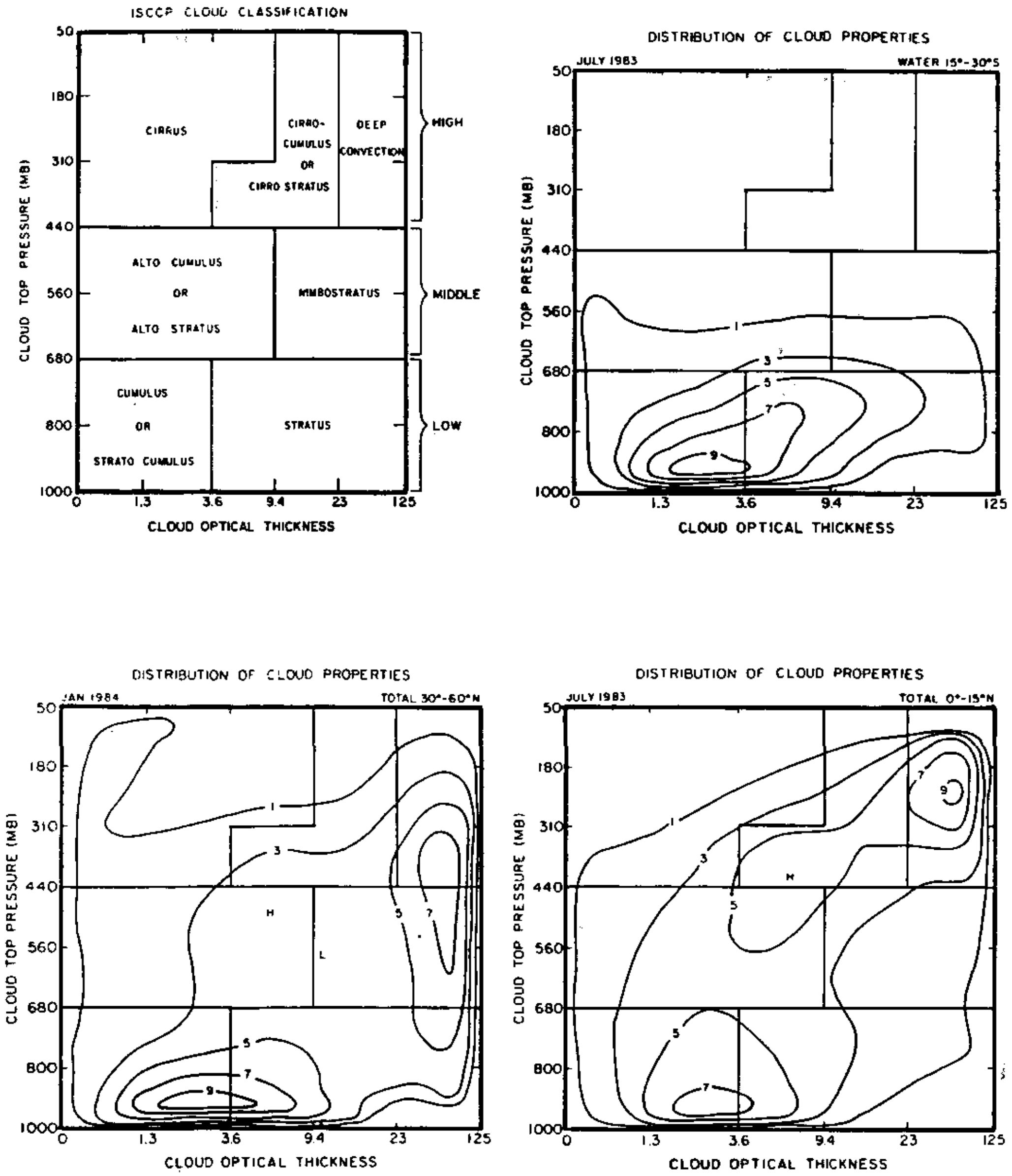


Figure 6.3: *Classification and frequencies of occurrence of combinations of cloud optical thickness and top pressure for different latitude zones from ISCCP.*

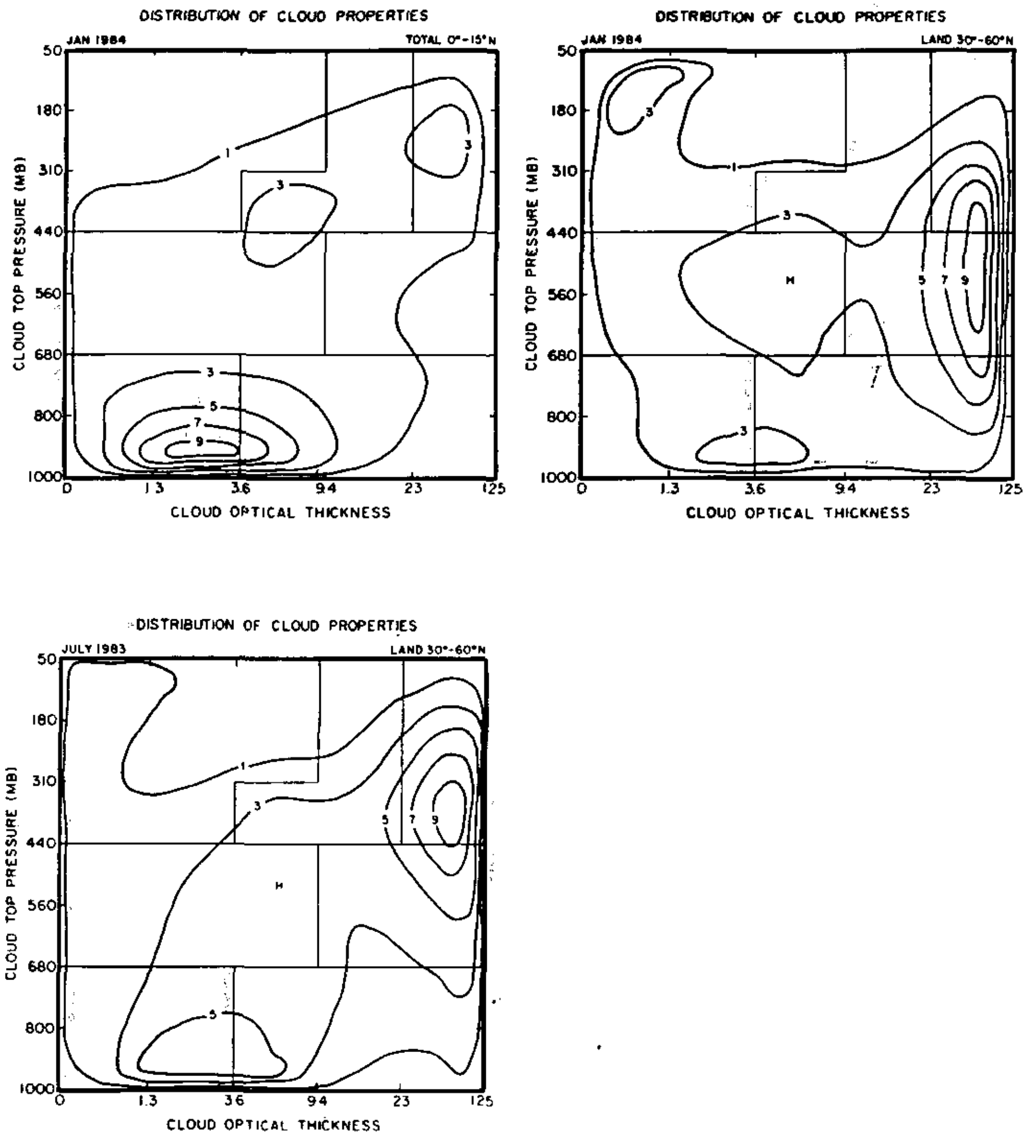


Figure 6.4: *Frequency of occurrence of different combinations of cloud optical thickness and top pressure from ISCCP.*

frequently (Fig. 6.3(d)). ITCZ cloudiness over land shifts latitude dramatically with season, generally occurring near the sub-solar latitude. Over ocean, on the other hand, there is usually evidence of a double structure (Fig. 6.1) that does not move, one part in each hemisphere; the summer part is much more active as evidenced by more frequent convective complexes and generally larger cloud cover. The dry (winter) season is characterized by more low-level, optically thin (fair weather) cloudiness (Fig. 6.4(a)).

At higher latitudes, the winter (colder) and summer (hotter) seasons differ in the relative frequency of occurrence of fair-weather and storm cloud types and in the average properties of the storm clouds. Winter brings more storm clouds (Fig. 6.4(b)). The winter storm clouds over oceans are generally optically thicker with higher cloud tops, (contrast Fig. 6.3(c) and 6.4(b)); over land cloud tops are higher in summer (Fig. 6.4(c)), but average optical thicknesses are larger in winter. Summer also brings somewhat more frequent fair-weather conditions, particularly over land where more cirrus clouds occur (Fig. 6.4(c)).

Although the seasonal temperature variations at low latitudes are quite small, the cloudiness changes are dramatic; whereas, the seasonal temperature changes at middle latitudes are very large, but the changes in cloud properties are more modest. The seasonal variations of cloud properties in these latitude zones are also opposite in sign: summer cloud amounts, optical thicknesses and top heights are larger in the tropics, while they are smaller in middle latitudes, as compared with winter values. In all cases, seasonal variations over land tend to be larger than over ocean. These facts all show that there are not simple relationships between cloud properties and surface climate, usually represented by temperature.

6.3 Diurnal variations

A similar variety of behavior is apparent in the style of daily cloud variations over land and ocean and at lower and higher latitudes. Higher latitude ocean clouds exhibit no significant diurnal variations. Winter clouds over land also show little variation; however, summer cloudiness has larger cloud cover and top heights, but lower optical thicknesses, in the afternoon (Fig. 6.5(a)). Subtropical latitudes show a weak tendency for increased cloudiness in the afternoon over land, but over oceans the maximum cloudiness occurs near dawn (Fig. 6.5(b)). The amplitude of this oceanic cycle is *significant despite the very small change in surface temperature it is associated with.* In the tropics, a variety of diurnal cycles over land and ocean, depending on the predominant cloud type and local meteorological conditions, combine to produce a more complicated cycle (Fig. 6.5(c)).

6.4 Why do clouds vary?

The names given to the radiative cloud types in Fig. 6.3(a) were selected to correspond, approximately, to the classical morphological cloud types identified by weather observers. A comparison of the frequencies of occurrence of these types shows that this correspondence, while not exact, is very good (Fig. 6.6). A relation between "dynamical" and "radiometric" cloud types is not unexpected, however, since it is the atmospheric motions that cause clouds: the more vigorous these motions are, the more cloud mass is expected to be produced. Going from the "fair-weather" cloud types (Fig. 6.6(b) and 6.6(c)) to the "stormy-weather" cloud types (Fig. 6.6(a) and 6.6(d)) corresponds to an increase in cloud top height and optical thickness (or water content). Even the cirrus amounts are similar (Fig. 6.6(e)).

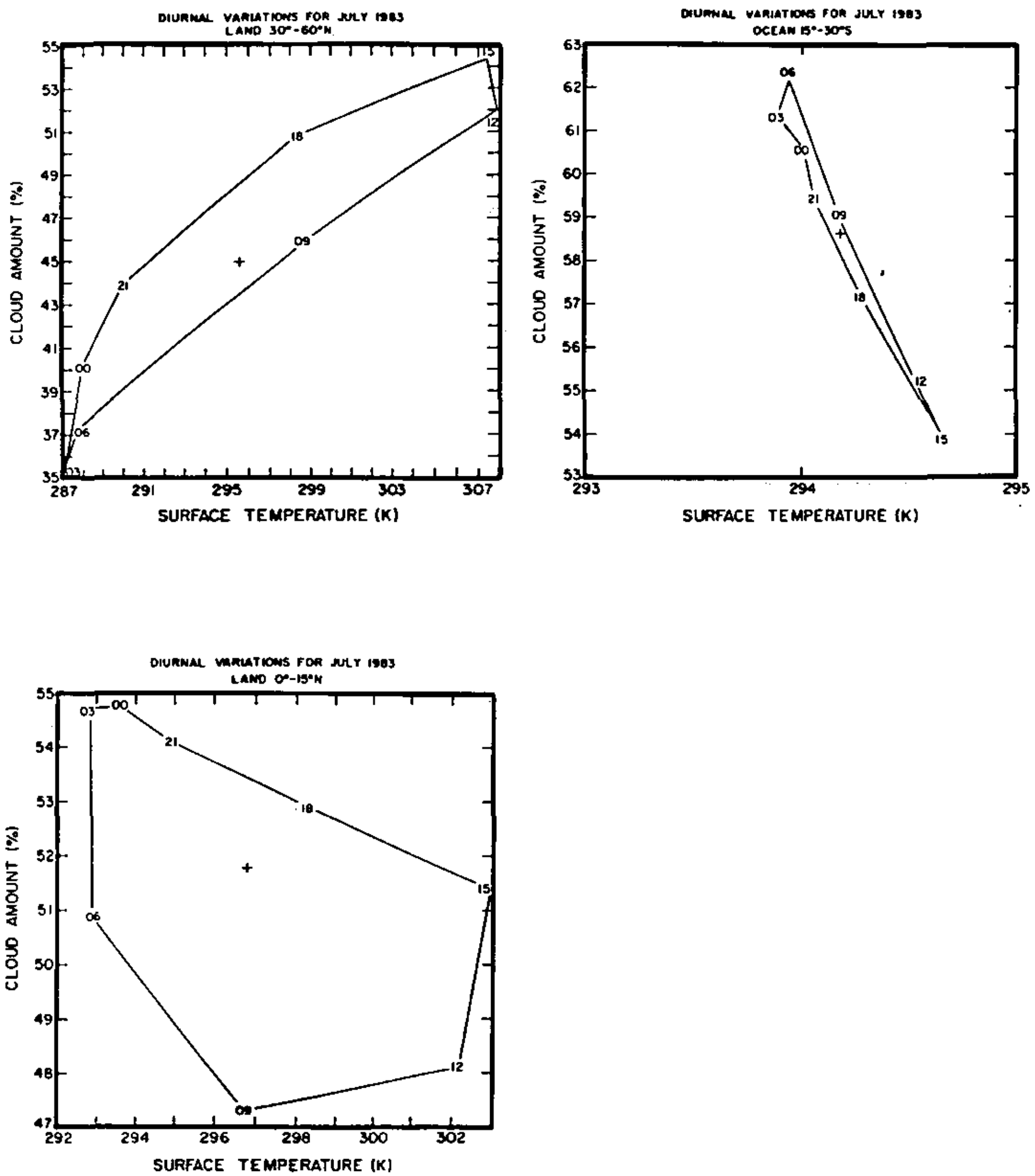


Figure 6.5: *Zonal mean cloud amounts (%) versus mean surface temperatures as a function of local time of day (0-24 hr) from ISCCP.*

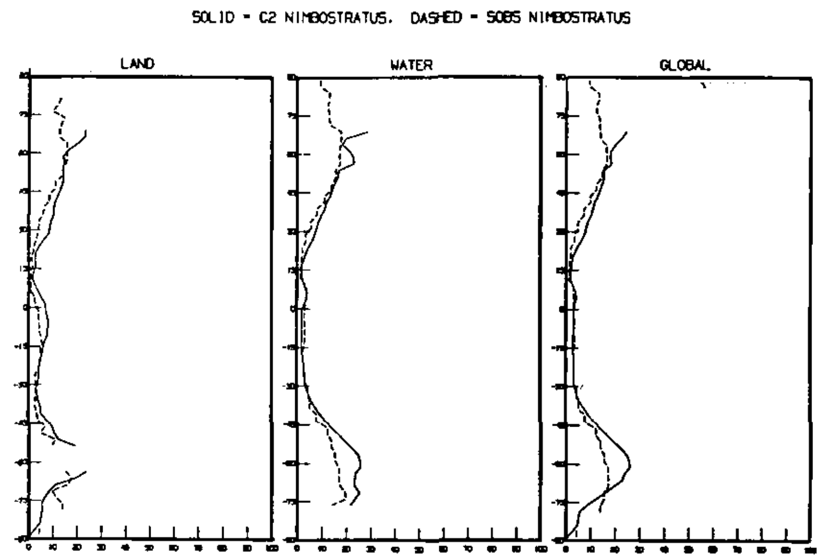


Figure 6.6: *(a) Zonal mean amount (%) of nimbostratus cloud in NH winter from surface observations and ISCCP.*

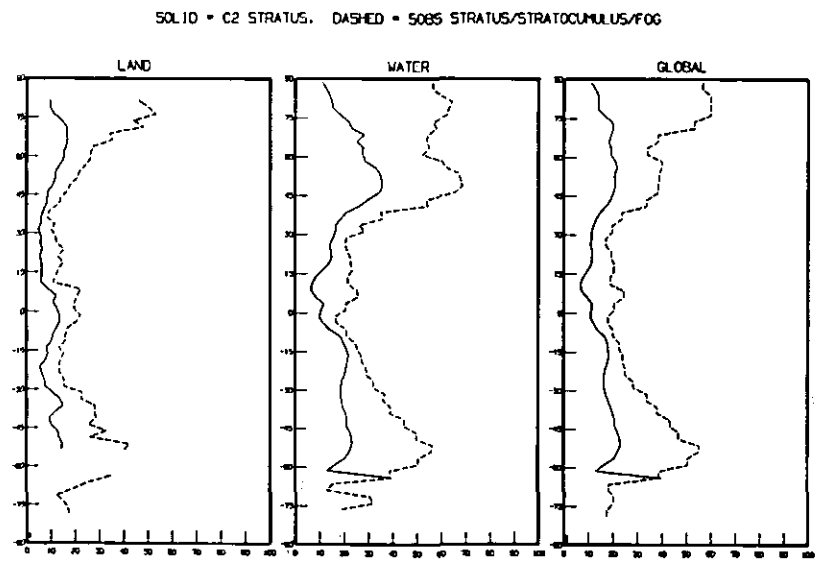


Figure 6.6: *(b) Zonal mean amount (%) of stratus cloud in NH summer from surface observations and ISCCP.*

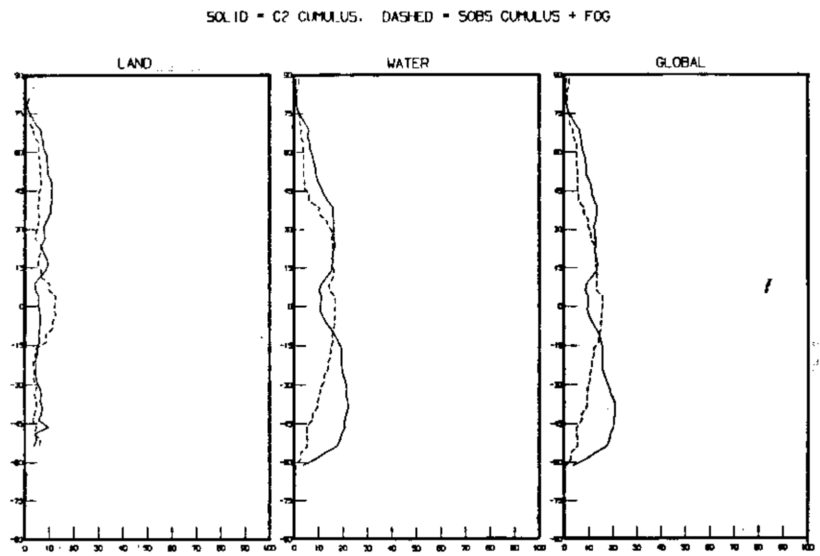


Figure 6.6: *(c) Zonal mean amount (%) of cumulus cloud in NH summer from surface observations and ISCCP.*

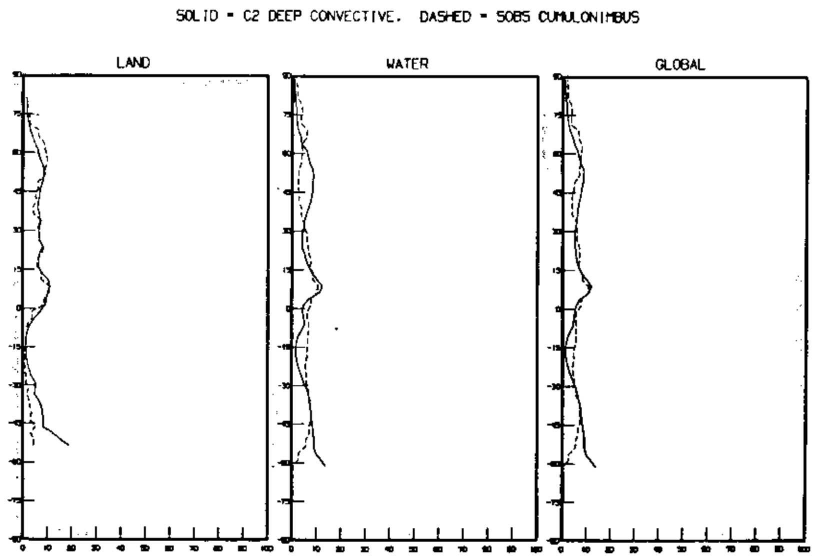


Figure 6.6: *(d) Zonal mean amount (%) of deep convective cloud in NH summer from surface observations and ISCCP.*

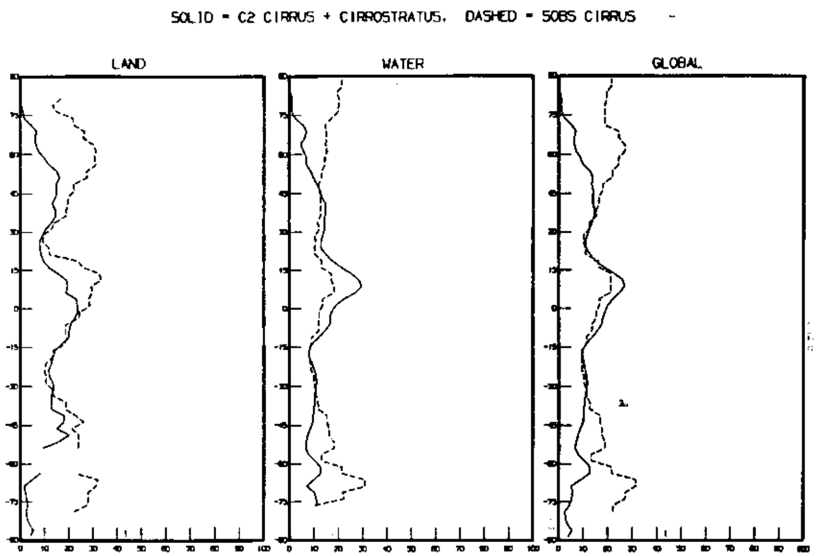


Figure 6.6: *(e) Zonal mean amount (%) of Cirrus cloud in NH autumn from surface observations and ISCCP.*

The results of several preliminary studies help elucidate this connection between cloud properties and atmospheric motions, but much more work is needed to quantify this physical relationship so that it can be used as the basis of improved cloud parameterizations in climate models. One such relationship is that between the amount of water formed in the cloud and the temperature at which the cloud forms. Since the condensation rate increases with temperature, it has been proposed that cloud water amount also increases (Somerville and Remer 1984; Betts and Harshvardhan 1984). However, this behavior also depends on how the precipitation rate changes with temperature, since it is the balance between condensation and precipitation that determines the cloud water amount. The observed behavior of the cloud optical thickness of low-level clouds (Fig. 6.7), which is proportional to water content when the vertical extent cannot change too much, changes sign between temperatures colder than about $-10\,^\circ C$ and warmer temperatures. At the colder temperatures, which occur mostly over land, (and where most of the available aircraft observations were collected), the cloud water amount tends to increase with increasing temperature; but at warmer temperatures, which compromise a majority of the Earth, cloud water amount decreases with increasing temperature. A possible explanation is that precipitation efficiency is higher at warmer temperatures (Tselioudis et al. 1992). This different perspective obtained from a global dataset illustrates one major advantage of satellite observations.

This explanation for the behavior in Fig. 6.7 is also consistent with the facts that the average cloud particle size and the temperature at which the sign of the cloud water - temperature relationship changes differ between land and ocean: the generally larger cloud particle sizes over ocean (Fig. 5.3 in Part I) are associated with a colder transition temperature as if precipitation is more efficient

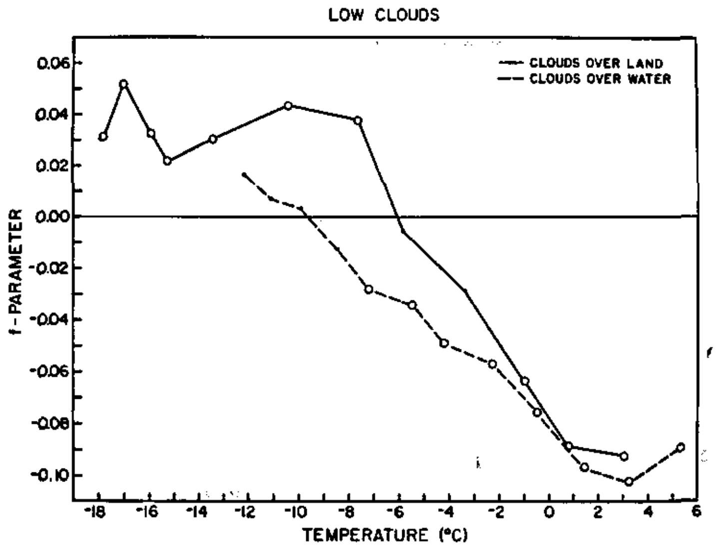


Figure 6.7: *Zonal mean values of $f = (1/\tau)(d\tau/dT)$ for low clouds over land and ocean from 1984 ISCCP results (after Tselioudis et al. 1991).*

in ocean clouds. Further study of the details of this relationship may help identify the key factors governing the formation of clouds and precipitation.

The location and structure of tropical convective complexes are the most highly variable cloud features on Earth. It has been proposed that sea surface temperature (SST) controls the spatial distribution of tropical convection based on observed variations with season and during El Niño events (e.g., Graham and Barnett 1987). Thus, a tight feedback between tropical SST and convection has been postulated, which also implies that the energy and water transports by convection, as well as the radiative flux changes associated with the convective cloud formations, are closely related to SST. A more detailed examination of the data (e.g., Fu et al. 1990) shows that the correlation between changes in SST and convective cloudiness is not actually very simple: Fig. 6.8 shows that the seasonal variations do not occur at the same time, that some areas with very warm surface waters can be free of convection and that some areas with cool surface waters are covered by convection. These results (Fu et al. 1990) show that this linkage cannot be considered as a simple one-dimensional relationship, but rather must be considered as three-dimensional (ie, dynamical), since the vertical structure of the atmosphere above the boundary layer as well as the character of the atmosphere in the boundary layer play equally important roles in determining the location of convective systems.

There is also some evidence that the sizes of convective complexes are related to the average properties of the clouds (Machado et al. 1992). As the systems grow larger, the cloud top heights increase and the average optical thicknesses increase (Fig. 6.9). This might be explained by all of these properties being determined by the magnitude of the updraft velocities in the convective towers. This supposition may also be supported by differences in lightning activity between ocean and land systems (Price and Rind 1991).

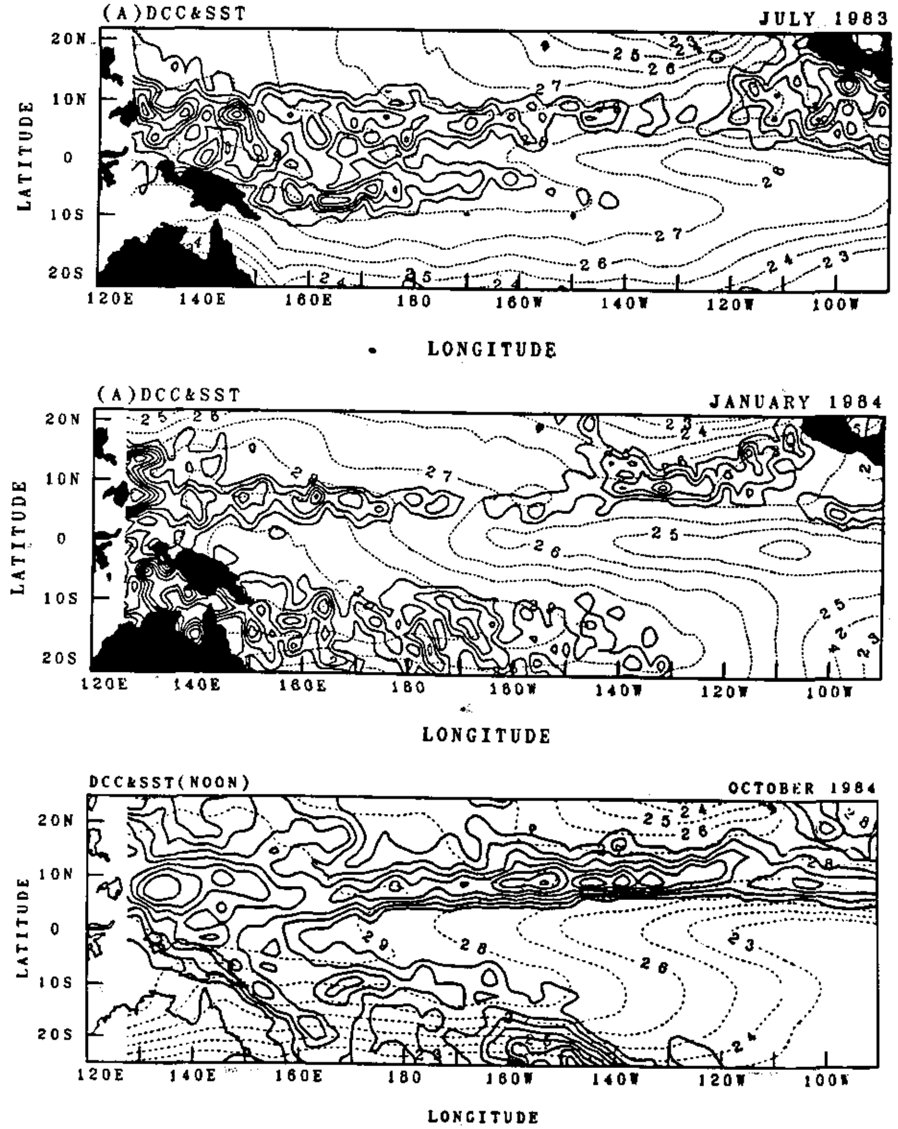


Figure 6.8: *Distribution of intense convection and sea surface temperatures in the tropical Pacific for three months (after Fu et al. 1990).*

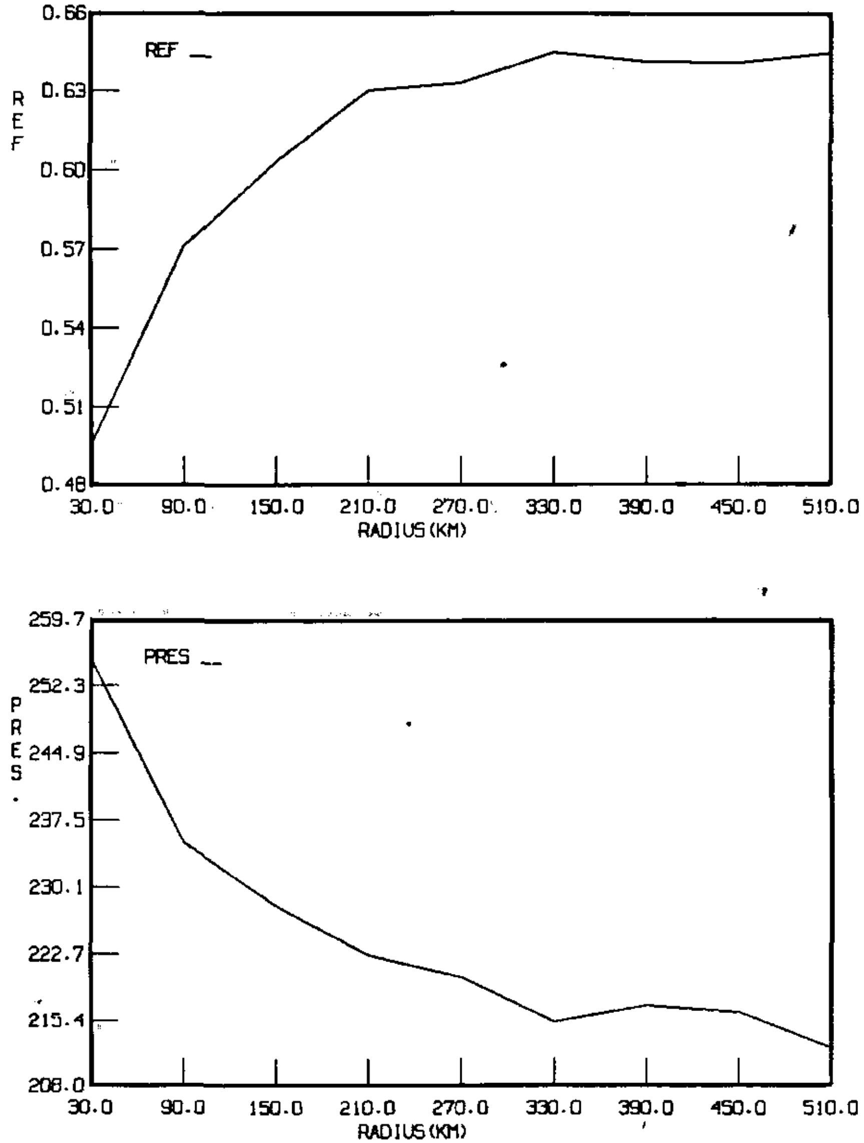


Figure 6.9: *Variation of average cloud reflectivity and top pressure with radius of the high (Temperature < 245 K) cloud clusters (based on Machado and Rossow 1992).*

6.5 Cloud effects on radiative flux variations

The basic radiation balance (Tab. 6.1) shows that over 70% of the solar flux absorbed by Earth is absorbed at the surface; however, the opacity of the atmosphere to terrestrial radiation allows little direct cooling of the surface by emitted radiation. This is the "greenhouse effect" which is due primarily to water vapor in the atmosphere. The net radiative heating of the surface is balanced primarily by the latent cooling associated with evaporation of water from the oceans (there is also some direct transfer of heat from the surface by turbulent atmospheric motions, particularly over land). The atmosphere is heated by the formation of precipitation, closing the water cycle, and emits this extra heat as terrestrial radiation. Thus, the net radiative balance of the atmosphere is a cooling.

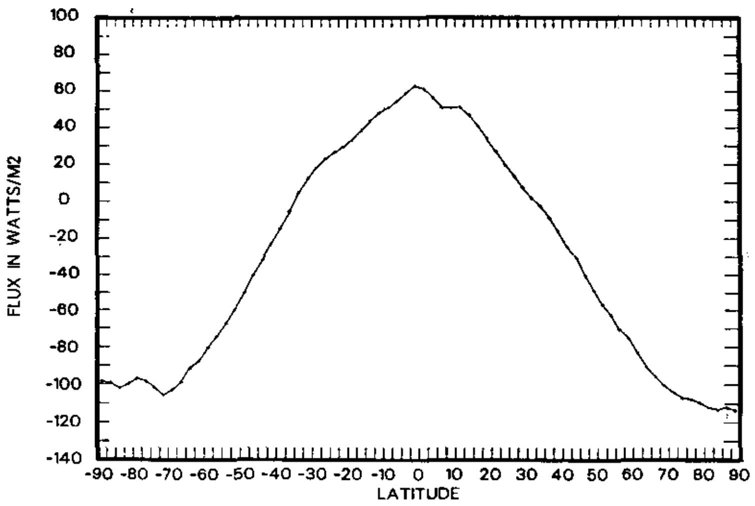
The addition of clouds to a planet with the same surface and atmospheric properties as Earth (with clouds) causes a decrease in the amount of solar radiation absorbed by the planet and a decrease in the amount of terrestrial radiation emitted by the planet (Tab. 6.1). Since the former change is twice a large as the latter, there is a net cooling effect (Harrison et al. 1990). If the imbalance caused by sudden removal of all clouds were allowed to continue, the surface temperature would rise. The increasing upward flux of terrestrial radiation from the surface would be absorbed by the atmosphere causing its temperature to rise. This purely radiative effect would probably be aided by increasing latent heat fluxes at higher temperatures. Eventually, as the surface and atmosphere approach a new, higher temperature equilibrium, the downward terrestrial flux from the atmosphere would nearly cancel the upward flux so that most of the surface cooling would be by latent (and sensible) heat transfer. Thus, the nature of the changes of energy exchanges in response to a perturbation away from equilibrium depends on the time scale considered: the most rapid response to changing surface temperature is caused by changes of the terrestrial radiation, but the final (new) balance would be determined by changes in latent and sensible cooling. In this sense, a rapid change in cloudiness results in a change of solar heating at the surface and a change of terrestrial heating of the atmosphere (cf., Stephens and Webster 1981).

Atmospheric and oceanic motions are driven by horizontal and vertical gradients in the heating. The Earth's spherical shape produces the primary contrast of solar heating between the equator and poles, but atmospheric motions transport some heat so that the net flux at the top of the atmosphere has the pattern shown in Fig. 6.10(a). Clouds alter the average latitudinal gradients of the radiative fluxes (Fig. 6.10(b)), serving to decrease the forcing for the general circulation in the tropics and near the poles and to increase it in midlatitudes.

Clouds also alter the vertical distribution of heating/cooling, because of their radiative effects and because of the latent heat exchanges associated with their formation (e.g., Stephens and Webster 1981). The nature of this effect depends on the amount water vapor in the atmosphere. However, we lack detailed enough information about cloud vertical structures to assess this effect quantitatively.

Small systematic variations of clouds over a diurnal cycle alter the radiative forcing that would obtain. The atmospheric response is more complicated than a simple local change in atmospheric turbulence. Over lower latitude oceans, since the surface temperature does not respond much to increased solar heating, it is the small change in atmospheric heating caused by the presence of clouds which controls

(a)



(b)

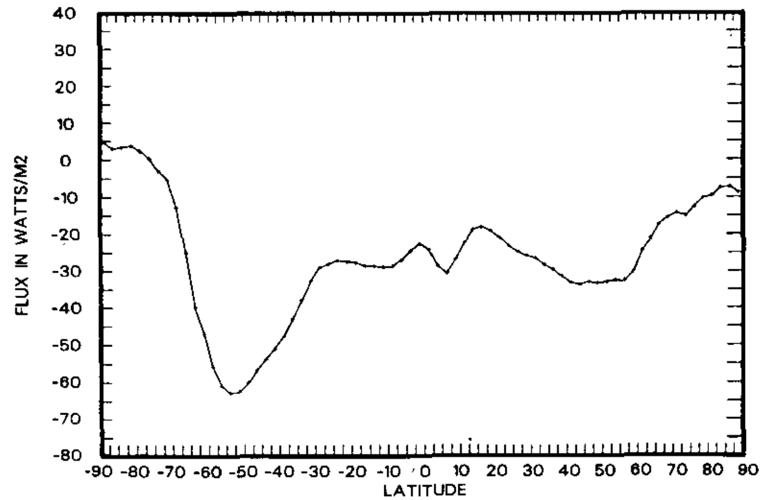


Figure 6.10: *(a) Zonal, annual mean net radiative flux at the top of the atmosphere, calculated from ISCCP datasets, (b) Net changes in the zonal, annual net radiative flux, shown in (a), caused by adding partial cloud cover.*

Quantity	ISCCP Values	
WITH CLOUDS		
Absorbed Solar at TOA	228.0	(232.4)
Emitted Terrestrial at TOA	232.4	
Net Radiation at TOA	-4.4	(0.0)
Absorbed Solar at SRF	162.2	(165.3)
Emitted Terrestrial at SRF	51.4	
Net Radiation at SRF	110.8	(113.9)
Absorbed Solar by ATM	65.8	(67.1)
Emitted Terrestrial by ATM	181.0	
Net Radiation by ATM	-115.2	(-113.9)
WITHOUT CLOUDS		
Absorbed Solar at TOA	280.2	(285.6)
Emitted Terrestrial at TOA	254.6	
Net Radiation at TOA	25.6	(31.0)
Absorbed Solar at SRF	212.7	(216.8)
Emitted Terrestrial at SRF	75.0	
Net Radiation at SRF	137.7	(141.8)
Absorbed Solar by ATM	67.5	(68.8)
Emitted Terrestrial by ATM	179.6	
Net Radiation by ATM	-112.1	(-110.8)
DIFFERENCE PRODUCED BY PARTIAL CLOUDINESS		
Absorbed Solar at TOA	-52.2	(-53.2)
Emitted Terrestrial at TOA	-22.2	
Net Radiation at TOA	-30.0	(-31.0)
Absorbed Solar at SRF	-50.5	(-51.5)
Emitted Terrestrial at SRF	-23.6	
Net Radiation at SRF	-26.9	(-27.9)
Absorbed Solar by ATM	-1.7	(-1.7)
Emitted Terrestrial by ATM	1.4	
Net Radiation by ATM	-3.1	(-3.1)
CONTRAST BETWEEN OVERCAST AND CLEAR		
Absorbed Solar at TOA	-68.6	(-69.9)
Emitted Terrestrial at TOA	-29.3	
Net Radiation at TOA	-39.3	(40.6)
Absorbed Solar at SRF	-67.0	(-68.3)
Emitted Terrestrial at SRF	-38.1	
Net Radiation at SRF	-28.9	(-30.2)
Absorbed Solar by ATM	-1.6	(-1.6)
Emitted Terrestrial by ATM	8.8	
Net Radiation by ATM	-10.4	(-10.4)

Table 6.1: *Global summary of annual average fluxes at the top of the atmosphere (TOA), at the surface (SRF) and in the atmosphere (ATM), with and without clouds, calculated from ISCCP datasets. All values are in Wm^{-2}. Values in parentheses are adjusted by requiring a flux balance at the top of the atmosphere in the annual mean. The third part of the table indicates the changes in the fluxes caused by removing (partial) cloud cover, while the fourth part indicates the differences between completely cloudy and completely clear conditions.*

the strength of boundary layer turbulence. Apparently, the daytime boundary layer is stabilized by heating in the clouds near the top of the layer; consequently, the cloud amount, optical thickness and height decrease throughout the day from a peak near dawn (e.g., Nicholls 1984). These changes shift the peak solar heating towards afternoon, but also shift the minimum terrestrial cooling towards morning. Since water vapor amounts are relatively high, the former effect is more important.

Over midlatitude land, the surface temperature responds quickly to the changing heating, so that it controls the boundary layer dynamics. In this case the larger temperatures near noon appear to enhance boundary layer turbulence and cloud amount, optical thickness and height all increase towards afternoon. Although these cloud changes serve to decrease the solar heating of the surface more rapidly than for clear conditions as sunset approaches, they also decrease the surface cooling, which may be a more important effect for the relatively dry boundary layer air over land.

The radiative effects of seasonal cloud variations are even more subtle. Figure 6.11(a) shows the seasonal extremes of the net radiation at the top of the atmosphere (cf., Fig. 6.10(a) for the annual means). At midlatitudes, cloud amount and optical thickness increase in winter, while cloud top temperature decreases primarily because the atmosphere is colder but also because the cloud top heights increase slightly. Increasing optical thickness and cloud amount enhance the wintertime decrease of solar heating caused by the changing inclination of Earth's rotation axis; but the colder cloud temperatures imply an offsetting decrease in cooling by terrestrial radiation. The clouds switch from a net cooling for summer to almost no net effect in winter (Fig. 6.11(b)). In the tropics, the cloud changes act in the opposite way: cloud amount, optical thickness and cloud height increase in summer. Increasing optical thickness and cloud amount decreases the summertime increase of solar heating; but the colder cloud temperatures imply an offsetting decrease in cooling by terrestrial radiation. These two effects appear to cancel, (Fig. 6.11(b)), leaving the net effect of the clouds about the same – a weak cooling. These changes appear to strengthen the mean atmospheric circulation by increasing the latitudinal contrast of radiative heating.

6.6 Summary - Possible cloud-climate feedbacks

A closer look at the results, assuming they are accurate enough, may suggest much more subtle cloud feedbacks (Rossow and Lacis 1990); however, the "observed" cloud radiative feedbacks on diurnal, seasonal and El Niño changes cannot be used directly to represent the feedbacks on climate. In all these cases that we can observe, we are monitoring the changes in cloud properties and radiative fluxes that occur during "rapid" deviations from equilibrium. They are rapid in the sense that the temperatures do not change much; hence the changes in fluxes that occur will depend on whether the temperatures have had time to change or not. As an example, the cloud effects shown in Table 6.1 are valid only for the case where the temperatures do not change and imply that both the solar and terrestrial net fluxes change most at the surface; but as the surface temperature changes, the largest terrestrial flux change caused by removing the clouds shifts from the surface to the atmosphere. A strong hint that the results may differ when the temperature changes is given by the completely different behavior exhibited by ocean and land areas: over oceans, perturbations which change the clouds produce changes in the radiation balance without significant changes in temperature, whereas over land, surface temperature changes occur almost immediately, producing a change in the terrestrial fluxes in addition to those produced by changing clouds. Thus, the actual radiative effect of a cloud change depends on the time scale on which it occurs. All of the observational

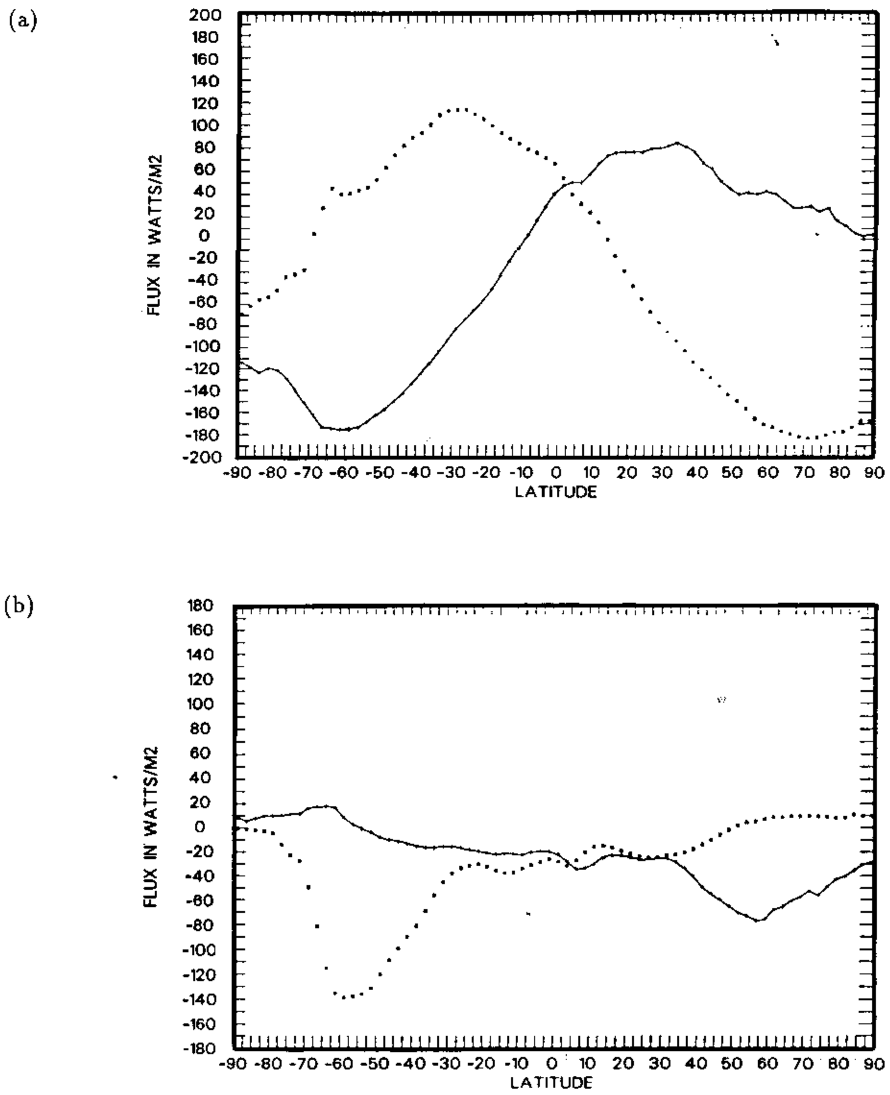


Figure 6.11: *(a) Zonal mean net radiative flux at the top of the atmosphere for January and July, calculated from ISCCP datasets, (b) Net changes in the zonal net radiative flux for January and July, shown in the upper figure, caused by adding partial cloud cover.*

evidence we have seems to suggest that cloud radiative feedbacks are negative for short-time scales (e.g., Hartmann and Short 1980; Rossow and Lacis 1990; Ardanuy et al. 1991), implying that clouds tend to suppress variations. What happens at longer time scales, we do not know as yet.

Despite the large amount of work that has been completed to produce the ISCCP, ERBE and NIMBUS-7 datasets, they are just the beginning of the more difficult studies needed to establish the relationships between atmospheric motions, cloud properties and their effects on radiation. By combining these datasets with other meteorological and surface datasets, some of these relationships can be found. Moreover, these data can be exploited to examine the role of clouds in the water cycle, too. This work is only just beginning.

There are also plenty of satellite (and other) datasets that have not been fully analyzed. In some cases the techniques exist but no one has undertaken the large analysis task. Examples of this kind are systematic analysis of satellite measurements of surface temperature and albedo and use of infrared sounder measurements to infer cloud properties. In other cases the analysis techniques still need to be developed. Examples of this are improvements in the infrared sounder methods and further development of microwave analysis methods to base them more on physical radiation models.

In the next 5-10 years, new kinds of satellite measurements are planned that also require development of new types of analysis methods. The most notable cases are:

1. methods for a simultaneous analysis of information from multiple instruments, each making measurements at many wavelengths, covering the electromagnetic spectrum from UV to microwave,
2. methods for extracting the full information from real spectrometers, and
3. analysis methods that can exploit the information contained in the polarization of reflected sunlight and scattered microwaves.

6.7 References

Ardanuy PE, Stowe LL, Gruber A, Weiss M (1991) Shortwave, longwave and net cloud-radiative forcing as determined from NIMBUS-7 observations. *J Geophys Res* 96: 18537-18549

Betts AK, Harshvardhan (1984) Thermodynamic constraint on the cloud liquid water feedback in climate models. *J Geophys Res* 92:8483-8485

Fu R, DelGenio AD, Rossow WB (1990) Behaviour of deep convective clouds in the tropical Pacific deduced from ISCCP radiances. *J Atmos Sci* 3:1129-1152

Graham N, Barnett TP (1987) Observations of sea surface temperature and convection over tropical oceans. *Science* 238:657-659

Harrison EF, Minnis P, Barkstrom BR, Ramanathan V, Cess RD, Gibson GG (1990) Seasonal variation of cloud radiative forcing derived from the Earth Radiation Budget Experiment. *J Geophys Res* 95:18687-18703

Hughes (1984) Global cloud climatologies: A historical review. *J Climate Appl Meteor* 23:724-751

Machado LAT, Rossow WB (1992) Structural characteristics and radiative properties of tropical cloud clusters. *Mon Wea Rev* (submitted)

Machado LAT, Desbois M, Duvel J-P (1991) Structural characteristics of deep convective systems over tropical Africa and Atlantic Ocean. *Mon Wea Rev* 120: 392-406

Nicholls S (1984) The dynamics of stratocumulus: Aircraft observations and comparisons with a mixed layer and model. *Quart J Roy Meteor Soc* 110:783-870

Price C, Rind D (1992) A simple lightning parameterization for calculating global lightning distributions. *J Geophys Res* 97: 9919-9933

Rossow WB, Lacis AA (1990) Global, seasonal cloud variations from satellite radiance measurements. Part II: Cloud properties and radiative effects. *J Climate* 3:1204-1253

Rossow WB, Schiffer RA (1991) ISCCP cloud data products. *Bull Amer Meteor Soc* 72:2-20

Somerville RCJ, Remer LA (1984) Cloud optical thickness feedbacks in the CO_2 climate problem. *J Geophys Res* 89:9668-9672

Stephens GL, Webster PJ (1981) Clouds and climate: Sensitivity of simple systems. *J Atmos Sci* 38:235-247

Tselioudis G, Rossow WB, Rind D (1992) Global patterns of cloud optical thickness variation with temperature. *J Climate* : in press

Wallace JM, Hobbs PV (1977) Atmospheric Science, An Introductory Survey. New York, *Academic Press*

Warren SG, Hahn CJ, London J, Chervin RM, Jenne RL (1986) Global distribution of total cloud and cloud type amounts over land. *NCAR Tech Note* TN-273+STR/DOE Tech. Rep. ER/60085-H1. 29 pp. + 200 maps

Warren SG, Hahn CJ, London J, Chervin RM, Jenne RL (1988) Global distribution of total cloud and cloud type amounts over the ocean. *NCAR Tech Note* TN-317+STR/DOE Tech. Rep. ER-0406. 42 pp. + 170 maps

6.8 Additional reading

– **Cloud Climatologies**

Hughes (1984) Global cloud climatologies: A historical review. *J Climate Appl Meteor* 23:724-751

Rossow WB, Schiffer RA (1991) ISCCP cloud data products. *Bull Amer Meteor Soc* 72:2-20

Stowe LL, Yeh HTM, Eck TF, Wellemeyer CG, Kyle HL, the NIMBUS-7 Cloud Data Processing Team (1989) NIMBUS-7 global cloud climatology. Part II: First year results. *J Climate* 2:671-709

Warren SG, Hahn CJ, London J, Chervin RM, Jenne RL (1986) Global distribution of total cloud and cloud type amounts over land. *NCAR Tech Note* TN-273+STR/DOE Tech. Rep. ER/60085-H1. 29 pp. + 200 maps

Warren SG, Hahn CJ, London J, Chervin RM, Jenne RL (1988) Global distribution of total cloud and cloud type amounts over the ocean. *NCAR Tech Note* TN-317+STR/DOE Tech. Rep. ER-0406. 42 pp. + 170 maps

– Cloud Radiative Feedbacks

Cess RD (1976) Climate change: An appraisal of atmospheric feedback mechanisms employing zonal climatology. *J Atmos Sci* 33:1831-1843

Hansen J, Lacis A, Rind D, Russell G, Stone P, Fung I, Ruedy R, Lerner J (1984) Climate sensitivity: Analysis of feedback mechanisms. Climate Processes and Climate Sensitivity. *Geophys Monogr Ser*, Vol. 29.

Harrison EF, Minnis P, Barkstrom BR, Ramanathan V, Cess RD, Gibson GG (1990) Seasonal variation of cloud radiative forcing derived from the Earth Radiation Budget Experiment. *J Geophys Res* 95:18687-18703

Hartmann DL, Short DA (1980) On the use of earth radiation budget statistics for studies of clouds and climate. *J Atmos Sci* 39:431-439

Hartmann DL, Doelling D (1991) On the net radiative effectiveness of clouds. *J Geophys Res* 96:1204-1253

Paltridge GW (1974) Global cloud cover and earth surface temperature. *J Atmos Sci* 31:1571-1576

Rossow WB, Lacis AA (1990c) Global, seasonal cloud variations from satellite radiance measurements. Part II: Cloud properties and radiative effects. *J Climate* 3:1204-1253

Schneider SW (1972) Cloudiness as a global climate feedback mechanism: The effects on the radiative balance and surface temperature of variations in cloudiness. *J Atmos Sci* 29:1413-1422

Stephens GL, Webster PJ (1981) Clouds and climate: Sensitivity of simple systems. *J Atmos Sci* 38:235-247

Stephens GL, Greenwald TJ (1991a) The Earth's radiation budget and its relation to atmospheric hydrology, 1, Observations of the clear sky greenhouse effect. *J Geophys Res* 96:15311-15324

Stephens GL, Greenwald TJ (1991b) The Earth's radiation budget and its relation to atmospheric hydrology, 2, Observations of cloud effects. *J Geophys Res* 96:15325-15340

Wang W-C, Rossow WB, Yao M-S, Wolfson M (1981) Climate sensitivity of a one-dimensional radiative-convective model with cloud feedback. *J Atmos Sci* 38:1167-1178