

CLOUDS AND THEIR EFFECTS ON THE CLIMATE SYSTEM

Covering the Earth on scales as large as millions of square kilometers, airborne water particles transport energy and affect the balance between incoming and outgoing radiation.

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The balance between incoming solar radiation and outgoing infrared radiation determines to a significant degree the Earth's climate. Clouds play a major role in determining the net radiative balance, so any change in cloud coverage or optical properties leads to a new climate state. To study the present and future climate states, the most comprehensive numerical tools currently available are global climate models. A major challenge in these models of Earth's environment is an accurate representation of clouds. (See figure 1.)

Clouds and atmospheric circulation

To guide our understanding of the role that clouds play in the climate system, we will use results from a three-dimensional general circulation model, the National Center for Atmospheric Research's Community Climate Model.¹ The model was forced with observed sea surface temperatures for a ten-year period, 1979–88, and the results were averaged over the entire period. Figure 2 shows the zonally averaged thermal structure of Earth's lower atmosphere. The warmest temperatures, around 300 K, are at the lowest latitudes, with temperatures decreasing toward the poles. In the vertical, the temperature decreases uniformly from the surface up to 10–15 kilometers. This region of decreasing temperature, the troposphere, is deepest in the tropics and shallowest at the poles. What maintains this thermal structure and what role do clouds play in its maintenance?

The radiative source of energy for the Earth's surface-atmosphere system is a balance between absorbed

solar shortwave radiation, with wavelengths between 0.2 and 4 microns, and outgoing longwave radiation emitted to space, with wavelengths between 4 microns and a few hundred microns (see figure 3). Figure 4 shows this balance $N(T)$ —the net incoming shortwave radiation minus outgoing longwave radiation—at the top of the atmosphere, as produced by the general circulation model. There is a net gain of energy between 40° N and 40° S, with a peak gain of 70 W/m² at the equator. Poleward of 40°, there is a net loss; hence the outgoing longwave emission to space exceeds the absorbed shortwave radiation. These results indicate why tropical temperatures exceed those at high latitudes.

However, a simple analysis that assumes the thermal structure is determined solely by radiative processes results in tropical temperatures too high by at least 30 K and polar temperatures too cold by 40 K. This indicates that the state of the atmosphere is not solely determined by radiative processes. Small-scale convective motions efficiently transport heat and moisture in the vertical, while large-scale horizontal atmospheric motions transport heat from the tropics to higher latitudes. These motions are driven by radiative, latent and sensible heat processes.

The atmospheric transport of dry static energy F_A or latent energy F_Q is described by the following two equations:

$$\nabla \cdot F_A = N(A) + LP + SH \quad (1)$$

$$\nabla \cdot F_Q = LH - LP \quad (2)$$

Equation 2 is the energy form of the mass balance for atmospheric water vapor. LP is the latent heat produced by condensation of water vapor, which forms clouds. SH is the surface sensible heat flux, and LH is the latent

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Tropical clouds in the Caribbean. Clouds not only create spectacular visual displays but also play an important role in determining Earth's climate. They may also be a major factor in determining how the climate responds to human influence. **Figure 1**

heat of evaporation from the surface into the atmosphere (see figure 3). $N(A)$ is the net radiative flux in the atmosphere:

$$N(A) = N(T) - N(S) \quad (3)$$

Here $N(S)$ is the net radiative flux into the Earth's surface. At the surface, energy (F_O) can also be transported in the oceans:

$$\nabla \cdot F_O = N(S) - LH - SH \quad (4)$$

Here the available energy to drive ocean heat transport is a balance between the incoming net radiative energy—shortwave minus longwave—and the loss from the ocean to the atmosphere due to latent and sensible heat transfer. The total transfer of heat for the atmosphere and ocean system is obtained by adding equations 1, 2 and 4:

$$\nabla \cdot [F_A + F_Q + F_O] = N(T) \quad (5)$$

It is important to note that the atmosphere and ocean systems are coupled through the relations given in equations 1–5. This article focuses on the atmospheric component, but we should always keep in mind that the surface and atmosphere should be treated together, as a system.

What role do clouds play in defining the magnitude and distribution of these terms that drive latitudinal energy transport? To answer this question, it is best to separate the radiative fluxes into clear and cloudy components. Thus we consider the net radiative flux,

$$N(A) = N_{\text{clr}}(A) + NCF(A) \quad (6)$$

where N_{clr} is the clear-sky net radiative flux and NCF is the net cloud radiative forcing. Figure 5a shows the annual zonal mean of $N(A)$ and N_{clr} from the general circulation model. Note that its magnitude is over 100

W/m^2 at most latitudes and that it is negative; N_{clr} is also large in magnitude and negative. Finally, figure 5b shows $NCF(A)$. This result indicates that clouds warm the atmosphere between the latitudes of 50° north and south, and cool the atmosphere poleward of 50° . The maximum warming is located near 5° N. As shown in figure 5b, $NCF(A)$ is dominated by the longwave cloud-forcing component, since present models of cloud properties indicate little shortwave absorption. However, as discussed below, these models may severely underestimate cloud shortwave absorption.

The large negative N_{clr} , and hence large negative $N(A)$, results from the significant penetration of shortwave flux through the atmosphere to the surface, which is viewed as a loss of energy for the atmosphere. The maximum clear-sky shortwave atmospheric absorption is 85 W/m^2 in the tropics, compared with an incident flux of 420 W/m^2 at the top of the atmosphere. In the longwave, the atmosphere emits radiation both to space and to Earth's surface. This emission obeys the Stefan-Boltzmann law and is proportional to T^4 . Thus, radiatively the atmosphere loses energy.

Of course, the atmosphere also transports energy in the form of latent heat, as described by equation 2, so the total transport of latent plus dry static energy, called moist static energy, is governed by

$$\nabla \cdot [F_A + F_Q] = N(A) + LH + SH \quad (7)$$

The right-hand side of this equation is shown in figure 6. We now see that the tropical cloud forcing, mainly longwave radiation of 30 W/m^2 , is roughly one-half of the total 60-W/m^2 forcing due to poleward transport of moist static energy. For this reason, changes in tropical longwave cloud forcing can lead to significant changes in large-scale circulation. A study using a general circulation model

investigated the role of longwave cloud forcing by removing it from the atmosphere.² Removing the atmospheric longwave cloud forcing reduces the strength of the large-scale circulation, the Hadley cell, by a factor of two. Thus, atmospheric longwave cloud forcing is an important component in forcing atmospheric circulation.

A similar analysis at the Earth's surface indicates that shortwave fluxes are the dominant factor in $N(S)$; hence according to equation 4, they play an important role in ocean heat transport. At the surface, shortwave cloud forcing dominates in the tropics, with a forcing of -60 to -70 W/m^2 , while in the extratropics, between 40° and 60° latitude, shortwave and longwave cloud forcing at the surface are equally important, roughly 20 – 50 W/m^2 . At high latitudes, longwave cloud forcing dominates, with a magnitude of 60 W/m^2 . In comparison, the latent heat flux from the surface ranges from 160 W/m^2 in the tropics to around 60 W/m^2 in the extratropics. These values indicate that surface shortwave cloud forcing is an important component of the total surface heat budget.

Through equation 5, the radiation balance $N(T)$ is defined as the *forcing* of the climate system. If a perturbation is made to the present climate system by increases in greenhouse gases, volcanic aerosols or solar variations, then the perturbed climate system will seek a new equilibrium state. Figure 7 shows the initial change in $N(T)$ due to increases in greenhouse-effect gases from the pre-industrial period to the present.³ This change is due to increases in CO_2 , CH_4 , N_2O and the chlorofluorocarbons. The latitudinal dependence of the change in $N(T)$ is dominated by the T^4 emission, which obeys the Stefan-Boltzmann relation. Note that the magnitude of this change appears quite small compared with that shown in figure 4; it is these apparently small changes in $N(T)$ that

lead to climate change. Cloud processes within the climate system can respond to the initial perturbation. The changes in these processes determine the *feedback* of the climate system.⁴ A quantitative measure of the feedback is defined by $[\Delta N(T)/\Delta T_s]^{-1}$, the so-called climate sensitivity parameter. Hence, climate forcing is determined by $N(T)$, while climate feedback is determined by the derivative of $N(T)$.

An example of a negative cloud feedback to the climate system involves low-level marine stratus clouds. Calculations indicate⁵ that a 14% increase in these clouds is sufficient to offset the initial forcing from a doubling of CO_2 . Because marine stratus clouds are so low in the atmosphere, the shortwave cooling dominates the longwave warming effect.

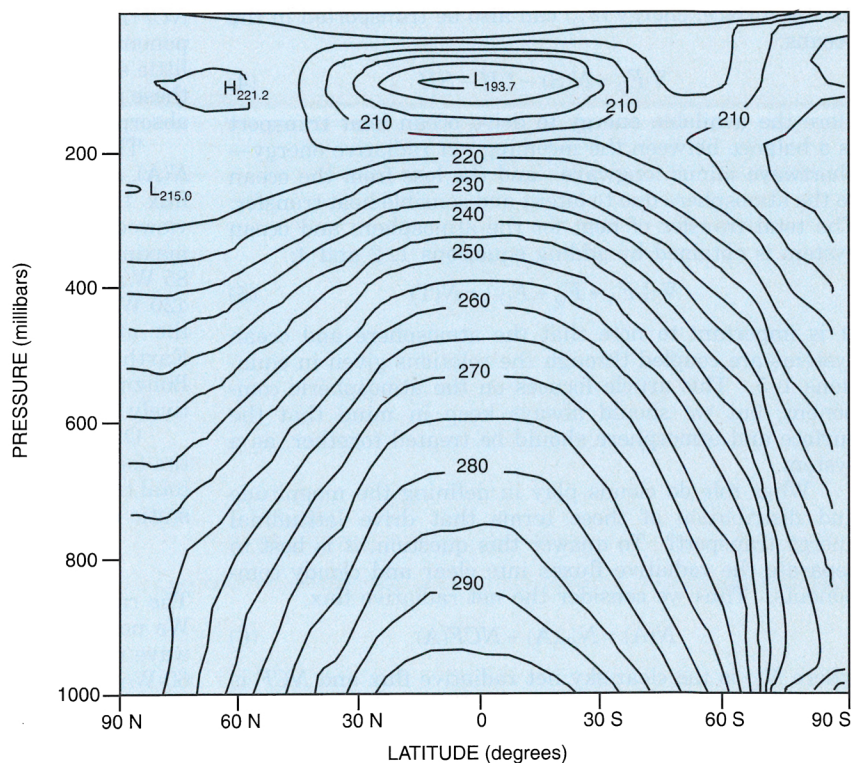
An example of a positive cloud feedback mechanism involves cloud altitude. If a warmer climate leads to an increase in cloud height, these higher, cooler clouds will decrease the outgoing longwave flux—that is, increase $N(T)$. For a 500-m increase in cloud height and an assumed upper-level cloud cover of 20%, the increase in $N(T)$ is 2.4 W/m^2 , which is as large as the tropical forcing due to the initial increase in trace gases shown in figure 7.

Clouds are important to the radiative and latent processes that control the poleward transport of energy of the present climate system. Cloud feedback can also significantly increase or decrease the initial perturbation of this system. To narrow the uncertainties in our capability to model these process, we must do further research on a number of cloud properties.

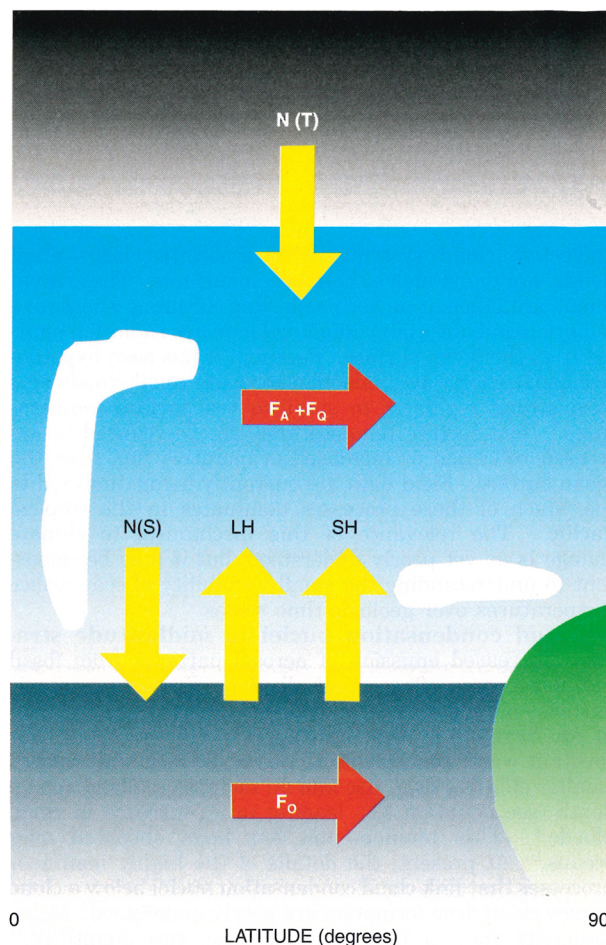
Cloud properties

A fundamental problem in modeling how clouds affect the climate system involves scale. Cloud droplets form on

Temperature structure of the atmosphere. Contours show the annual zonal mean atmospheric temperature (in kelvin) from a 10-year integration of the National Center for Atmospheric Research's Community Climate Model. The contour interval is 5 K. **Figure 2**



Energy transport in the atmosphere. The schematic diagram shows the vertical fluxes of radiative energy at the top of the atmosphere $N(T)$ and at Earth's surface $N(S)$; the fluxes of latent heat LH and sensible heat SH ; and the horizontal transport of moist static energy $F_A + F_Q$ and ocean heat F_O . **Figure 3**



aerosol particles that have dimensions of a few hundredths to a few tenths of a micrometer. Cloud drops grow to tens of microns in diameter, while cloud coverage can reach over 1000 kilometers in diameter in the tropics. The time scales range from tens of seconds for microphysical processes to a day for the life cycle of cloud systems.

Thus the cloud properties that affect the radiative energy budget span 14 orders of magnitude in space and more than 3 orders of magnitude in time. Current numerical climate models can resolve spatial scales on the order of a few hundred kilometers and time scales on the order of tens of minutes. Thus there is considerable space and time disparity between the physical processes that generate and maintain clouds and the scales that are resolvable in climate models.

The only solution to the problem of including cloud effects in climate models has been to represent these detailed physical processes parametrically, a technique known as parameterization. It is important to note that parameterizations should incorporate the fundamental physics of the processes being modeled. The challenge in much of cloud-climate research is to use observation, theory and numerical modeling to understand and parameterize clouds for climate models, for it is only in the context of a global climate model that all of the important physical processes of the climate system can be included.

Relevant cloud microphysical properties

Clouds are composed of water particles in the liquid or solid phase. Cloud droplets form on aerosol particles in the atmosphere, which act as sites for water vapor condensation. The efficiency with which an aerosol particle can nucleate (initiate growth of) a cloud drop depends on both its size and its solubility. Condensation continues until the particles are sufficiently large (20 microns) to begin collisional interaction, which leads to even larger cloud droplets. At any given time and location, the number of drops at each particular size is described by the distribution $n(r)$. Liquid cloud droplets are spherical in shape. Ice particles, however, are in general nonspherical. Accounting for the effect of the nonspherical shape on the transfer of shortwave and longwave radiation is an important problem in cloud-climate research.

If the particles are spherical and their refractive index is known, then one can use solutions to Maxwell's equations to calculate the efficiency with which they scatter and absorb radiation.⁶ These solutions yield three fundamental quantities that define the optical properties of clouds: the extinction optical depth τ_{ext} , the phase function $P(\Omega, \Omega')$ (where Ω defines the solid angle for incident radiation and Ω' defines the solid angle of scattered radiation) and the single-scattering albedo ω_0 . The extinction optical depth describes the efficiency with which radiation is scattered or absorbed by cloud droplets; for shortwave radiation, it is proportional to the total droplet projected area:

$$\tau_{\text{ext}} \propto \int_0^{\infty} n(r) r^2 dr \quad (8)$$

The phase function defines the probability of a photons

being scattered into a unit solid angle. Typically, the larger the particle, the more likely a photon will be scattered in the direction of the incident photon (the "forward" direction). The single-scattering albedo measures the fraction of radiation scattered relative to the total particle extinction.

Detailed information about the distribution $n(r)$ is not necessary when the particles are relatively large compared with the wavelength of radiation. The cloud optical properties then depend upon two bulk microphysical quantities, the cloud liquid-water concentration

$$LWC = \frac{4}{3}\pi\rho_l \int_0^{\infty} n(r) r^3 dr \quad (9)$$

and the effective drop size

$$r_e = \frac{\int_0^{\infty} n(r) r^3 dr}{\int_0^{\infty} n(r) r^2 dr} \quad (10)$$

By combining equations 8–10, we find

$$\tau_{\text{ext}} \propto \frac{LWC}{r_e} \quad (11)$$

Cloud-climate issues

Three issues are particularly important to understanding the effect of clouds on climate:

▷ **Tropical anvils.** The warmest ocean surface temperatures occur in the tropical western Pacific. Deep

convective cloud systems, with associated large anvil clouds, are found above these ocean surfaces. These anvil clouds are efficient at both reflecting incoming shortwave radiation and absorbing outgoing longwave radiation emitted from the ocean surface. Recently, it has been hypothesized that shortwave reflection by these anvil clouds acts as a strong regulator to stabilize sea surface temperatures.⁷ Sea surface temperatures can be reduced by increased reflection or increased evaporative flux from the ocean surface. Field data are currently being analyzed to see which of these processes dominates in the tropical Pacific. The relevance of this mechanism to climate change is as yet poorly understood, but it may be important to understanding the small variability of sea surface temperatures over geologic time scales.

▷ **Cloud condensation nuclei in midlatitude stratus.** Increased emission of aerosol particles from fossil fuel usage may affect cloud albedos. These hygroscopic particles may increase cloud condensation nuclei and lead to an increase in the number of cloud droplets. For constant water concentration, there will be a shift toward smaller effective drop size and an increase in cloud optical depth (see equation 11), leading to an increase in cloud albedo.⁸ This phenomenon has been observed over oceans.⁹ At present, the details of the highly nonlinear processes that link cloud condensation nuclei below a cloud to new cloud drop formation are poorly understood. Measurements show a link between these two quantities.¹⁰ Also, the assumption that the liquid-water concentration remains fixed may not hold in general. The problem of linking increased aerosols to increased cloud albedo is under theoretical, modeling and observational study.

▷ **Cloud absorption.** Although cloud reflection is the major determinant of shortwave cloud forcing, shortwave absorption is non-negligible. Two recent observational studies indicate that the amount of shortwave absorption may be much larger than previous models suggest and that this effect is a global phenomenon.^{11,12} These studies indicate that shortwave cloud absorption in the atmo-

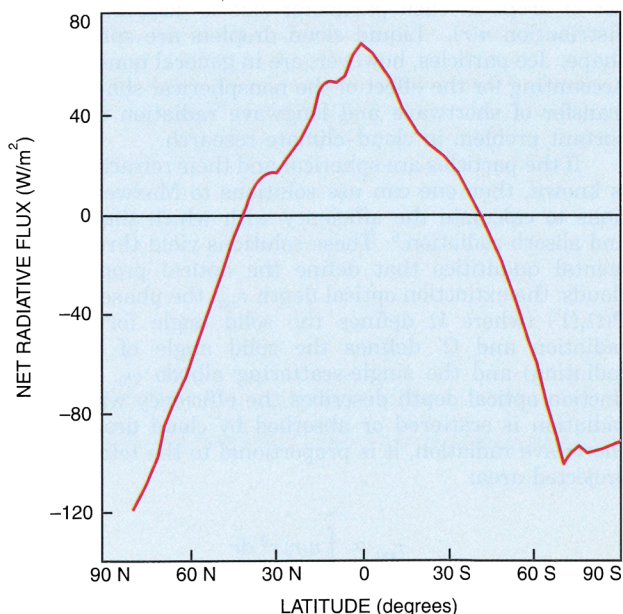
sphere is about 50% of the shortwave cloud forcing at the top of the atmosphere. This result implies that the absorption is nearly 40 W/m^2 , instead of the 2 W/m^2 indicated in our energy budget analysis. If these studies are confirmed, the $NCF(A)$ term in our energy budget analysis becomes a major determinant of the poleward heat transfer in equation 7.

Observation, modeling, theory

Approaches to understanding cloud-climate problems have focused on observation, numerical modeling and theory. Unlike experimental physics, observations of clouds must occur in an open, uncontrolled system—namely, the Earth's atmosphere.

Observation. Progress on many of the present cloud-climate problems is limited by sparse observational data. Observations of cloud radiation processes can be made from three locations: the Earth's surface, *in situ* and at the top of the atmosphere, via satellite remote sensing. The advantage of ground-based observations is that instruments can be calibrated and maintained over a long time period, enabling long-term observations of clouds in a number of locations; new instruments to complement existing ones can be introduced at any time. The DOE Atmospheric Radiation Measurements program is an example of a long-term ground-based program to collect data on cloud radiative processes.¹³ The limitation of ground-based measurements is that certain cloud characteristics, such as microphysical properties, cannot be observed from the ground, although some information can be retrieved by remote sensing from ground-based instruments, such as microwave sensing of column-cloud water. Also, there are many regions, such as the ocean, where it is logistically difficult to base surface instruments.

In situ observations from airplanes or balloons are



Net radiative flux at the top of the atmosphere $N(T)$. Plotted is the annual and zonally averaged net radiative flux, which is equal to the shortwave absorbed flux minus the outgoing longwave flux. **Figure 4**

Flux and cloud forcing. **a:** Annual and zonal mean net radiative flux into the atmosphere $N(A)$ and clear-sky component $N_{\text{clr}}(A)$. **b:** Net cloud forcing $NCF(A)$ and its shortwave and longwave components. **Figure 5**

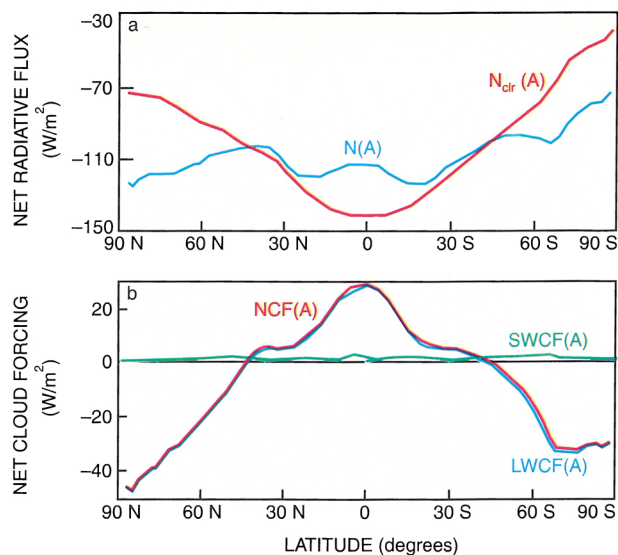
valuable in obtaining data on microphysical processes in clouds. For example, *in situ* instruments are needed to measure the cloud drop-size distribution $n(r)$. Aircraft instruments can also measure the radiative fluxes at the tops and bottoms of clouds, and even within a cloud. Thus these types of measurements are best suited to studying detailed microphysical processes occurring in or near a cloud system. The limitation of this approach is the difficulty in obtaining a sufficiently large data base for climate parameterization. The data are also not comprehensive, since payload and power limitations restrict the number of instruments on a given airplane. However, *in situ* data can be used to test satellite retrieval methods and microphysical models. The cost of *in situ* measurements is often prohibitive. Although *in situ* observations are important for understanding a given cloud process, they are limited in direct applications to climate research.

Satellites in space provide near-global coverage. Earth-radiation-budget instruments have measured fluxes on scales of roughly 30 kilometers, while other instruments, such as the Advanced High-Resolution Radiometer, can measure radiance at a specific wavelength down to a scale of about 1 km. The smallest scales observable from space using Landsat imagery are in the range of tens of meters. Hence, satellite data span a wide range of spatial scales.

Earth radiation-budget information, which measures the net radiative flux at the top of the atmosphere $N(T)$, is extremely important for global models. Every global model must be validated against accurate Earth radiation-budget data. NASA's Earth Radiation Budget Experiment, which ran from October 1985 to April 1989, provided calibrated data (see the article by V. Ramanathan, Bruce Barkstrom and Edwin Harrison in *PHYSICS TODAY*, May 1989, page 22), but at present there are no observations being taken of the Earth radiation budget. Because Earth radiation-budget data yield little information about vertical distribution of radiant energy, other satellite instruments, such as radiometers aboard geostationary satellites, must be used in conjunction with radiation models. Interpreting such measurements is difficult, however, because of the need to make assumptions about cloud properties that are still poorly understood. To improve these retrieval methods requires coupling satellite observations with *in situ* and ground-based observations. Other limitations to satellite methods are cost, payload and power restrictions. NASA is planning to launch a suite of instruments for cloud research as a part of the Earth Observing System.

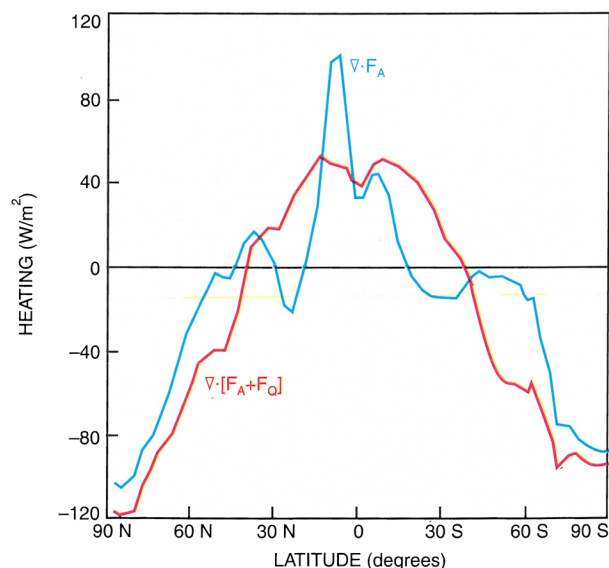
What is required for the near term is to use existing satellite instruments for cloud studies. The calibration of many current satellite instruments can be carried out with the help of *in situ* instruments. Such calibration is perhaps the highest-priority need for climate researchers.

Modeling. Numerical modeling on various scales of cloud systems is integral to a coherent research program. Present computational resources are able to represent an ensemble of clouds with scales ranging from 1 km up to 1000 km. Numerical simulations of resolved cloud systems can routinely be made in two dimensions (horizontal

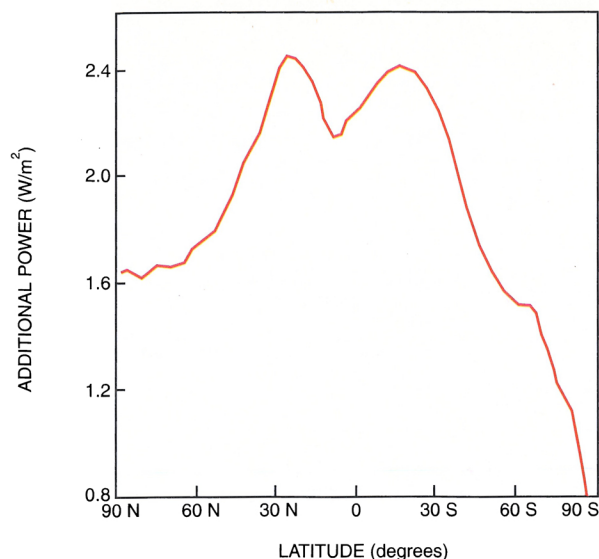


and vertical slices) and, in more limited cases, can be carried out in three dimensions. These models typically rely on imposed boundary conditions (such as large-scale horizontal winds) to represent the large-scale forcing from the climate system (with scales of motion greater than 1000 km). Future attempts should be made to allow for the interaction between the cloud ensemble and the large scale. Thus the smaller-scale cloud model can be forced with large-scale atmospheric conditions predicted by a climate model.

Global models need to be tested more quantitatively. At present, validation of general circulation models usually employs monthly mean satellite data on the Earth's radiation budget. Although this is important, models should be tested against time-varying data as well. Comparison



Atmospheric poleward heat transport of dry static energy (blue curve) and moist static energy (red curve). (See figure 3.) F_A represents dry static energy; F_Q represents latent energy. **Figure 6**



Greenhouse forcing. Plotted is the change in the net top-of-atmosphere radiative forcing $N(T)$ due to increases in greenhouse gases— CO_2 , CH_4 , N_2O , CFC_{11} and CFC_{12} —from preindustrial times to the present. **Figure 7**

of models with observations on time scales both longer and shorter than monthly means is required.

For example, tropical sea surface temperatures exhibit a variation every two to three years called El Niño. This variation of 2–3 °C in sea surface temperature forces changes in cloud cover and associated radiative fluxes. The response of models forced with the observed El Niño sea surface temperature patterns can be tested against Earth radiation-budget data.

Models should also be tested on shorter time scales, such as the diurnal scale. Comparison of modeled cloud properties with three-hourly data from the International Satellite Cloud Climatology Program are of great value.¹⁴

Theory. Theoretical research is needed on the fundamental questions of scale. How are the multiple cloud scales related? How does this multitude of scales affect the transfer of radiation within clouds? How can the relevant smaller-scale processes be incorporated into a general circulation model cloud parameterization? One focus of research on these questions is the study of the fractal nature of clouds.¹⁵ But further effort is required to relate the current studies to the climate parameterization problem. Present-day general circulation models assume that clouds are geometrically plane-parallel and thus neglect radiation entering or exiting from their sides. Also, horizontal cloud–cloud radiative interactions are neglected. Monte Carlo techniques reveal significant differences in cloud radiative properties obtained from an array of finite clouds and from a plane-parallel cloud.¹⁶

For parameterization purposes, the question is, How do we include these finite cloud radiative effects over scales of hundreds of kilometers? Theoretical studies of cloud geometry and cloud inhomogeneity are needed, again with special emphasis on relating these theoretical findings to the large-scale (on the order of a few hundred kilometers) climate problem. Further theoretical understanding of microphysical processes is needed, in particular, the role of aerosols on cloud drop-size distributions. Finally, given the extreme lack of understanding of tropi-

cal anvils, especially with regard to the microphysical processes, further theoretical work is desperately needed.

Observation, modeling and theory cannot occur in isolation from one another. Continued progress in understanding how clouds affect the climate system relies on an active interchange among these activities.

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