

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.

Jeffrey T. Kiehl



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



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.

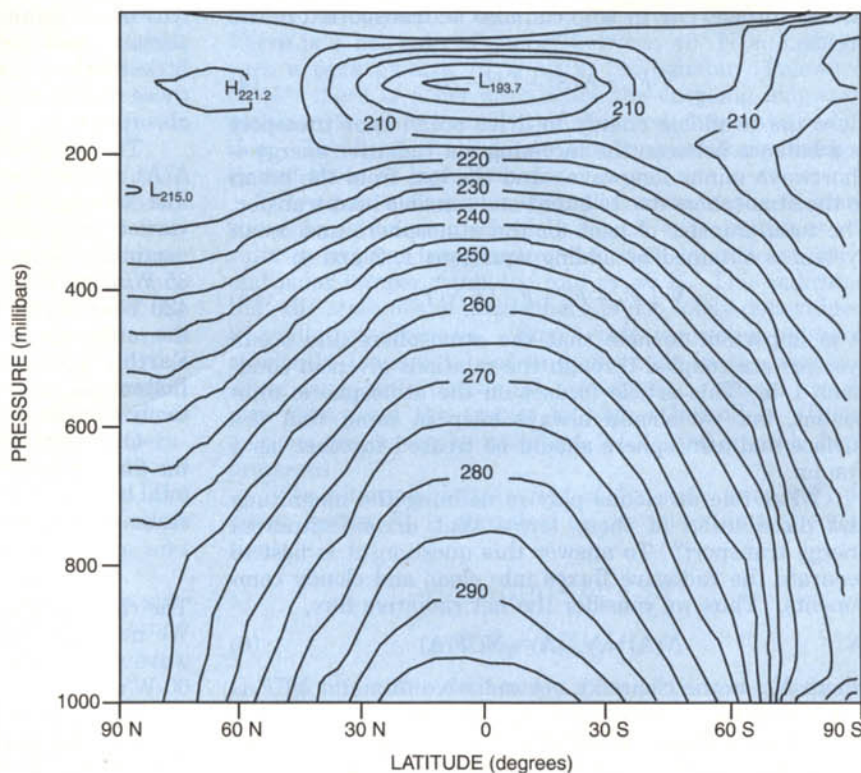
An example of a positive cloud feedback mechanism involves cloud altitude. If a warmer climate leads to an increase in cloud height, these higher, cooler clouds will decrease the outgoing longwave flux—that is, increase $N(T)$. For a 500-m increase in cloud height and an assumed upper-level cloud cover 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 processes, we must do further research on a number of cloud properties.

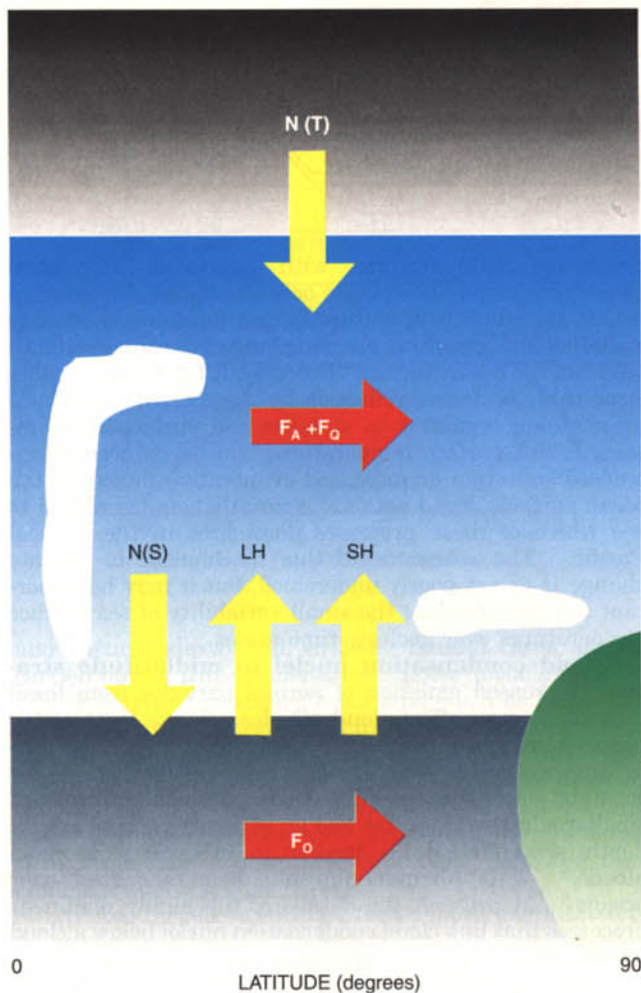
Cloud properties

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