

REVIEW ARTICLE

Cloud Feedbacks in the Climate System: A Critical Review

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ABSTRACT

This paper offers a critical review of the topic of cloud–climate feedbacks and exposes some of the underlying reasons for the inherent lack of understanding of these feedbacks and why progress might be expected on this important climate problem in the coming decade. Although many processes and related parameters come under the influence of clouds, it is argued that atmospheric processes fundamentally govern the cloud feedbacks via the relationship between the atmospheric circulations, cloudiness, and the radiative and latent heating of the atmosphere. It is also shown how perturbations to the atmospheric radiation budget that are induced by cloud changes in response to climate forcing dictate the eventual response of the global-mean hydrological cycle of the climate model to climate forcing. This suggests that cloud feedbacks are likely to control the bulk precipitation efficiency and associated responses of the planet’s hydrological cycle to climate radiative forcings.

The paper provides a brief overview of the effects of clouds on the radiation budget of the earth–atmosphere system and a review of cloud feedbacks as they have been defined in simple systems, one being a system in radiative–convective equilibrium (RCE) and others relating to simple feedback ideas that regulate tropical SSTs. The systems perspective is reviewed as it has served as the basis for most feedback analyses. What emerges is the importance of being clear about the definition of the system. It is shown how different assumptions about the system produce very different conclusions about the magnitude and sign of feedbacks. Much more diligence is called for in terms of defining the system and justifying assumptions. In principle, there is also neither any theoretical basis to justify the system that defines feedbacks in terms of global–time–mean changes in surface temperature nor is there any compelling empirical evidence to do so. The lack of maturity of feedback analysis methods also suggests that progress in understanding climate feedback will require development of alternative methods of analysis.

It has been argued that, in view of the complex nature of the climate system, and the cumbersome problems encountered in diagnosing feedbacks, understanding cloud feedback will be gleaned neither from observations nor proved from simple theoretical argument alone. The blueprint for progress must follow a more arduous path that requires a carefully orchestrated and systematic combination of model and observations. Models provide the tool for diagnosing processes and quantifying feedbacks while observations provide the essential test of the model’s credibility in representing these processes. While GCM climate and NWP models represent the most complete description of all the interactions between the processes that presumably establish the main cloud feedbacks, the weak link in the use of these models lies in the cloud parameterization imbedded in them. Aspects of these parameterizations remain worrisome, containing levels of empiricism and assumptions that are hard to evaluate with current global observations. Clearly observationally based methods for evaluating cloud parameterizations are an important element in the road map to progress.

Although progress in understanding the cloud feedback problem has been slow and confused by past analysis, there are legitimate reasons outlined in the paper that give hope for real progress in the future.

1. Introduction

In his 1905 correspondence to C. G. Abbott, T. C. Chamberlain notes

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“Water vapor, confessedly the greatest thermal absorbent in the atmosphere, is dependent on temperature for its amount, and if another agent, as CO₂, not so dependent, raises the temperature of the surface, it calls into function a certain amount of water vapor which further absorbs heat, raises the temperature and calls forth for more vapor. . . .”

This comment provides an early yet clear perspective on the problem of water vapor feedback (e.g., Held and

Soden 2000). Projecting the salient points of this comment onto the topic of cloud feedback exposes a number of important issues that highlight both the essential differences between cloud and water vapor climate feedback processes and touch on why progress in understanding cloud feedback has been particularly elusive. These differences are the essential topics of this paper.

- 1) The “thermal absorbent” character of water is greatly enhanced when in a condensed phase. On a molecule-by-molecule basis, water in either solid or liquid form in the atmosphere absorbs more than 1000 times more strongly¹ than in gaseous form. The amount of energy that is actually absorbed and scattered by clouds, however, depends on other complicating factors, many of which are relevant to cloud feedback.
- 2) It is reasonably clear what is meant by water vapor by Chamberlain’s remarks but it is much less obvious what is meant by clouds. As we will see, a number of cloud properties, including cloud amount, cloud height and vertical profile, optical depth, liquid and ice water contents, and particle sizes affect the radiation budget of earth and are all potentially relevant to the cloud feedback problem. This more extensive list of cloud parameters compared to water vapor merely reflects the greater complexity of the cloud feedback problem and explains why many different types of cloud feedbacks have been hypothesized. Wielicki et al. (1995) surveys a number of the feedback concepts proposed over the years.
- 3) From the system’s perspective of feedback discussed below, feedbacks require a definition of system output. In general, and in the context of Chamberlain’s comments specifically, this output is most commonly posed as global-mean (surface) temperature. In this context, the dependence of global-mean cloudiness on global-mean temperature is unclear and certainly much less obvious than for water vapor. To first order, clouds are governed more so by the large-scale motions of the atmosphere and thus on the complex dependence of the latter on the three-dimensional distribution of temperature. This complexity is arguably the single most important factor to cloud feedback yet is generally overlooked or grossly simplified in most reported cloud feedback studies.
- 4) Unlike water vapor, clouds also impart an almost equal effect on the disposition of solar radiation primarily through reflection (the so-called albedo effect). This introduces further complexity to the problem as, when viewed from the top of the atmo-

sphere (TOA), the thermal absorbent effects are largely offset by the albedo effect. The general tendency for long- and shortwave processes to broadly balance one another has confounded our attempts to understand the cloud feedback problem. In the 1970s, the observation that clouds produce a relatively small net effect on TOA fluxes was mistakenly interpreted as indicative of a negligible cloud feedback.

- 5) Clouds and water vapor are essential stages in the cycling of water between the earth and atmosphere. Clouds act both as sources and sinks of water vapor and water vapor is fundamental to the formation of clouds. Clouds also affect the earth system in a variety of other ways and couple many processes together over a wide range of time and space scales, a point that has also been understood for some time (e.g., Arakawa 1975). Thus it seems most unlikely that the feedbacks that involve clouds will operate independently of other feedbacks. Not only do clouds affect the strength of water vapor feedback but also they affect the strengths of snow/ice albedo feedbacks, soil moisture feedbacks (Betts 2000, 2004), and lapse rate feedbacks (e.g., Zhang et al. 1994) to mention a few. It is also argued in section 6 that cloud feedbacks govern the response of the global-scale hydrological cycle to climate forcings. The coupled nature of cloud feedback complicates our definition of the climate system and thwarts simple attempts to quantify cloud feedback effects on climate change.

Despite progress in many areas of the “cloud feedback problem,” progress has been slow, partly because of our vague concepts of feedback. This paper begins with a critical discussion on the nature of the feedback problem as usually posed in terms of the global-mean “climate sensitivity” and proposes that the main source of confusion over analysis of feedbacks lies not with the definition of feedback per se but with the definition of the system itself. It is also suggested that although the cloud–climate feedback problem is in principle complex for reasons mentioned, it is argued that there are a few essential processes that fundamentally characterize the problem. One of these concerns the governing effects of atmospheric motions, chiefly large scale, that organize global cloudiness. A second concerns the interaction of radiation and clouds and a brief overview of the effects of clouds on the radiation budget of the earth–atmosphere system is provided in section 3. Most of the discussion on this topic has focused on TOA radiative fluxes. The importance of clouds on the radiative budget of the atmosphere, mostly overlooked in feedback studies, is stressed. Cloud feedbacks in simple radiative–convective equilibrium (RCE) systems are then reviewed since these studies laid the early foundation for thinking about the topic. Three different classes of feedback concepts that attempt to explain the regula-

¹ The magnitude of this enhanced absorption can be simply inferred from the relationships between the broadband clear-sky emissivity and water vapor path and the equivalent broadband cloud emissivity and cloud liquid (or ice) water path.

tion of tropical SSTs are then reviewed in section 5. Section 6 returns to the systems perspective of the feedback problem and brings to view a number of issues that have limited the conclusiveness of these earlier and simpler cloud feedback studies and further exposes problems with the way system analysis is generally applied to analyze cloud feedbacks. Section 7 moves toward the more complex (model) climate systems as represented by climate general circulation models (GCMs) and briefly overviews the general problem of cloud parameterization in GCMs. Section 8 then describes a selection of GCM studies that highlight both the model sensitivity to cloud parameterizations and the effects of cloud feedbacks in GCMs. It becomes apparent that the feedback diagnostic tools used to analyze GCM experiments, generally rooted to systems analysis, are problematic and immature thus underscoring the need for new diagnostic approaches to study feedback. A central theme of the paper is returned to in section 9, which emphasizes the value for more stringent evaluation of cloud processes and their representation in GCMs. The final section summarizes the main themes of the paper

and offers an outlook for progress in the coming decade.

2. The nature of the problem: A critical discussion

Because of the profound influence of clouds on both the water balance of the atmosphere and the earth's radiation budget, small cloud variations can alter the climate response associated with changes in greenhouse gases, anthropogenic aerosols, or other factors associated with global change. Predictions of global warming by GCMs forced with prescribed increases of atmospheric CO_2 are uncertain, and the range of uncertainty has, seemingly, not changed much from initial estimates given decades ago. The effects of potential changes in cloudiness as a key factor in the problem of climate change has been recognized since at least the 1970s (e.g., Arakawa 1975; Schneider 1972; Charney 1979; among others). This point too is reiterated in Fig. 1 showing the range of surface warming estimates from a number of models that participated in the Coupled

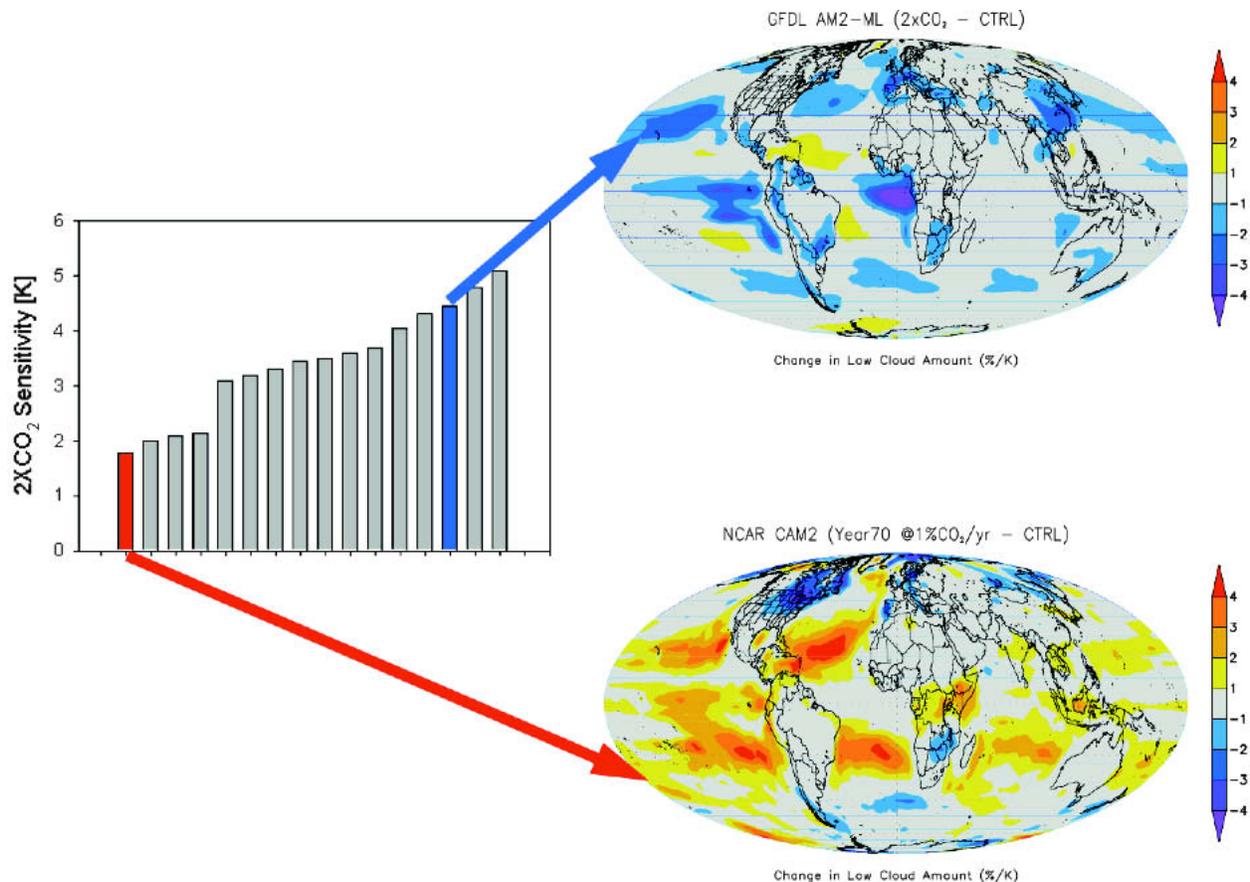


FIG. 1. The response of a number of present-day climate models forced by a $1\% \text{ yr}^{-1}$ increase of CO_2 . Shown is the difference of the 20-yr average of the simulation with fixed and increasing CO_2 . The averages are over years 1961–80; corresponding broadly to the time of a CO_2 doubling. To the right are the changes to low clouds averaged over this same period for two models that fall on either end of the projected warming range (courtesy of B. Soden).

Model Intercomparison Project (CMIP; Meehl et al. 2000; and more below).

Simple ideas taken from the control system theory are commonly used in an attempt to understand the different net responses of climate models to an imposed forcing. Much more on this topic is discussed in section 6, but for now we consider two different representations of climate as shown in Fig. 2. The top part portrays the view most commonly held when analyzing model data like that shown in Fig. 1. The bottom part portrays a more complex and perhaps more realistic system that operates on an entirely different time and space scale. The system portrayed at the top is meant to represent the global-time-mean climate, its mean input, and a measure of output usually thought of in terms of the difference between two equilibrium systems—one forced and one unforced. This is certainly an attractive view of a complex system and has value when comparing different model responses to the same forcing (e.g., Fig. 1). A more physical justification for these global-mean responses follow from elementary global energy balance considerations assuming small perturbations to this balance and linear, global responses. Given these considerations, it can be shown that (e.g., North et al. 1981)

$$\Delta T_s = \Delta Q / \lambda, \quad (1)$$

where we interpret ΔT_s as the global-mean (surface) temperature change (e.g., Figs. 1 and 2) and ΔQ is the

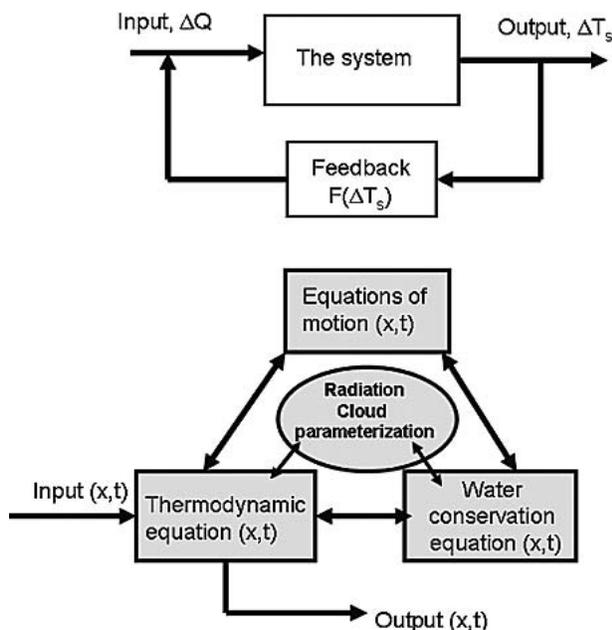


FIG. 2. (top) The common, simple system view of the global-mean climate system with feedbacks. (bottom) A view of one component of the climate system expanded for detail indicating the set of connecting subcomponents, each composed of processes that couple to other processes in space (x) and time (t).

radiative forcing that induces this change. Since most models impose similar values of ΔQ (at least for prescribed increases of CO_2), the response of the model to the forcing (i.e., ΔT_s), merely reflects the model sensitivity λ . As Fig. 1 indicates, there is currently a large discrepancy in the value of λ derived from different models. This discrepancy is widely believed to be due to uncertainties in cloud feedbacks (e.g., Webster and Stephens 1984; Cess et al. 1990; Senior and Mitchell 1993; Houghton et al. 1995, and in section 8). This point too is implied in Fig. 1 showing the changes in low clouds predicted by two versions of models that lie at either end of the range of warming responses. The reduced warming predicted by one model is a consequence of *increased* low cloudiness in that model whereas the enhanced warming of the other model can be traced to *decreased* low cloudiness.

The relationship between global-mean radiative forcing and global-mean climate response (temperature) is of intrinsic interest in its own right. A number of recent studies, for example, discuss some of the broad limitations of (1) and describe procedures for using it to estimate ΔQ from GCM experiments (Hansen et al. 1997; Joshi et al. 2003; Gregory et al. 2004) and even procedures for estimating λ from observations (Gregory et al. 2002). While we cannot necessarily dismiss the value of (1) and related interpretation out of hand, the global response, as will become apparent in section 9, is the accumulated result of complex regional responses that appear to be controlled by more local-scale processes that vary in space and time. If we are to assume gross time-space averages to represent the effects of these processes, then the assumptions inherent to (1) certainly require a much more careful level of justification than has been given. At this time it is unclear as to the specific value of a global-mean sensitivity λ as a measure of feedback other than providing a compact and convenient measure of model-to-model differences to a fixed climate forcing (e.g., Fig. 1).

It is tempting nevertheless to use the simple framework embodied in (1) to quantify model feedbacks and much has been written on this topic over the past 20 years. However, the real climate system connects different subcomponents, each evolving in time and each variable in space making the climate system more complex than the simplified view of it represented in the top part of Fig. 2. Although simple energy balance theory provides some framework for relating global-mean energetics to global-mean temperature, there is no clear theory that translates from the complex process-oriented system in Fig. 2 to the simpler global-mean system. Thus we have no clear theory that suggests the accumulated effects of cloud feedbacks are in any way a function of global-mean temperature or, as posed, ΔT_s . As we will see, the (usually unstated) assumptions about the nature of the system and its feedbacks, and how feedback processes relate specifically to surface temperature, dictate almost entirely the quantitative re-

sults from climate feedback analysis. This is alarming as we will see later how different assumptions about the system, applied to the same model output, produce feedback measures that not only differ in magnitude but also in sign.

Although there is presently not an obvious, correct approach to the analysis of feedbacks, there are some important lessons to be learned from the past feedback studies reviewed below. One message is there is an important need to evaluate critically the tools we use to quantify feedbacks, to encourage the development of new approaches to feedback analysis, to compare different methods of analysis, and disregard those deemed inappropriate. For example, Aires and Rossow (2003) propose a more general nonlinear multivariate approach to calculate the instantaneous sensitivities in contrast to the usual approach that calculates these sensitivities as differences in equilibrium states.

It is also the contention expressed in this paper that the inherent shorter time and smaller space scales of cloud processes fundamentally control how feedbacks take place. As such, the traditional global-mean perspective, while of some value, distorts our view of feedbacks and confuses our attempts to quantify them. One of the major omissions in mapping from a complex system defined on the intrinsic time and space scales at which the processes take place to its steady-state global-mean analog is the loss of the influence of the large-scale atmospheric circulation on clouds. It will be argued throughout this paper that one step toward unravelling the complex nature of cloud feedback lies ultimately in understanding such influences. It is the atmospheric circulation that broadly determines where and when clouds form and how they evolve. Cloud influences, in turn, feed back on the atmospheric circulation through their effects on surface and atmospheric heating, the latter involving a complex combination of radiative and latent heating processes, which, on the global scale, are intimately coupled (as discussed in sections 6 and 8). Therefore the basis for understanding this important feedback, in part, lies in developing a clearer understanding of the association between atmospheric circulation regimes and the cloudiness that characterizes these “weather” regimes (Fig. 3).

3. Effects of clouds on the radiation budget

Most cloud–climate feedback studies are concerned with processes associated with the transfer of radiation through and within clouds. Much of the focus of past feedback studies is concerned with the effects of clouds on the radiation balance at the TOA. This focus, in part, has been motivated by satellite experiments like the Earth Radiation Budget Experiment (ERBE) and the Clouds and the Earth’s Radiant Energy System (CERES) that provide quantitative measures of the instantaneous effects of clouds on the TOA radiation balance. Furthermore, the International Satellite Cloud

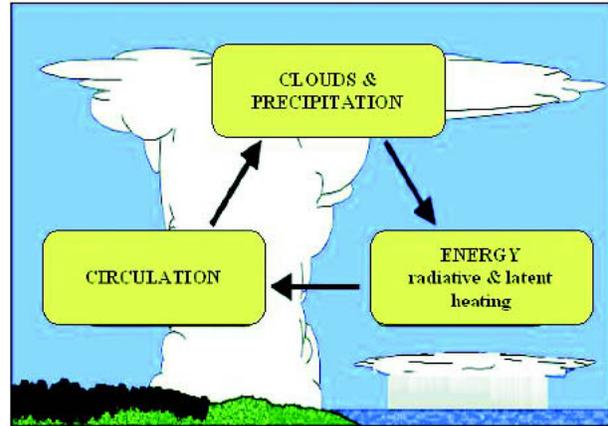


FIG. 3. A schematic depiction of the main elements of the cloud feedback problem. The links between the boxes indicate processes that establish the feedbacks.

Climatology Program (ISCCP) offers global distribution of total cloud cover and optical properties. The ISCCP formally began in 1983 with the collection of the first internationally coordinated satellite visible and infrared radiance data (e.g., Rossow and Schiffer 1991, 1999).

Satellite measurements of the type mentioned above are an important source of information for testing models especially when related to TOA fluxes and other properties of the climate system. These measurements offer much less information for understanding the separate effects of clouds on the atmospheric (ATM) and the surface (SFC) radiation budgets and it will be stressed how these influences are critical aspects of the cloud feedback problem. Much of what we know about the effects of clouds on the surface and atmospheric radiation budget is gleaned from model simulations directly or models constrained by certain types of data (e.g., Pinker and Corio 1984; Zhang et al. 1995; Charlack and Alberta 1996; and others). It will also be demonstrated how available TOA observations of clouds and TOA radiative fluxes cannot constrain critical assumptions about the relation between cloud physical and radiative processes—relationships of central relevance to the cloud feedback problem.

a. Effects of clouds on TOA radiation budget

In contemplating the effects of clouds on these different components of the energy balance, it proves convenient to consider differences between all-sky-measured fluxes and clear-sky fluxes in an effort to isolate the specific effect of cloud. This idea was introduced by Ellis and VonderHaar (1976) and popularized in the 1980s with analysis of TOA flux data collected as part of ERBE (Harrison et al. 1990). These all-sky–clear-sky flux differences can be interpreted in the following way. Suppose that

$$F_{\text{obs}} = (1 - N)F_{\text{clr}} + NF_{\text{cld}} \quad (2)$$

represents the flux observed in a cloudy scene, where N is the fractional cloud amount and F_{cld} is the flux associated with overcast portion of the scene. This expression is only an empirical approximation to a clear-cloudy-scene-averaged flux. It ignores cloud–cloud radiative transport that can occur in cumulus cloud fields for example producing nonlinear relationships between fluxes and cloud amount (e.g., Wienman and Harshvardhan 1984).

Simple rearrangement of (2) defines the quantity

$$C_{\text{LW}} = F_{\text{clr,LW}} - F_{\text{obs,LW}} = N[F_{\text{clr}} - F_{\text{cld}}(\varepsilon, T_c)], \quad (3a)$$

for longwave (LW) fluxes where the flux of the cloudy-sky portion of the sky is now noted to be a function of cloud (top) temperature T_c and emissivity ε . The analogous quantity for shortwave (SW) fluxes is

$$C_{\text{SW}} = F_{\text{clr,SW}} - F_{\text{obs,SW}} = N \frac{S_{\odot}}{4} (\alpha_{\text{clr}} - \alpha_{\text{cld}}), \quad (3b)$$

which is expressed in terms of the clear- and cloudy-sky albedos, α_{clr} and α_{cld} , respectively, and the TOA incident flux $S_{\odot}/4$. Although a misnomer, the quantities C_{LW} and C_{SW} continue to be referred to as cloud radiative forcing and, at least for TOA, it is possible to infer these quantities from global Earth Radiation Budget (ERB) observations. The common reference to these flux difference quantities as a forcing is confusing especially given that some studies suggest these quantities are in fact a measure of cloud feedback. In reality, these quantities are nothing more than a measure of the effect of clouds on the radiative budget relative to the clear-sky radiative budget. Hereafter these quantities are referred to as either the cloud flux effect or cloud flux.

The main point of (2) and (3) is that it provides a simple way of identifying cloud properties important to the radiative balance of the planet. For instance, we can infer that changes in both C_{LW} and C_{SW} arise from macroscopic changes of cloud cover N and/or the height of clouds (i.e., cloud temperature T_c). Changes in the albedo and emittance of the cloud also affect C_{SW} and C_{LW} , respectively, and these occur through bulk changes in cloud optical depth arising from variations in thickness and/or changes in cloud physical properties including particle size, water and ice contents, among others. Thus the range of parameters of possible relevance to the cloud feedback problem is potentially considerably larger than was considered in the very early feedback studies that were posed largely in terms of cloud top T_c and cloud amount N (e.g., Schneider 1972).

It was also stressed above how the net flux effects occur largely as a result of compensating effects on solar and infrared fluxes. This compensation can be expressed quantitatively in terms of the net cloud flux effect

$$C_{\text{net}} = C_{\text{LW}} + C_{\text{SW}}, \quad (4)$$

where generally, $C_{\text{LW}} > 0$, $C_{\text{SW}} < 0$, and the net response occurs typically as a small residual of these two competing effects. Here C_{LW} is broadly a measure of the greenhouse effect of clouds; the highest, coldest clouds that occur with tropical deep convective systems over the warmest sea surface temperatures (SSTs) induce the largest greenhouse effect (Fig. 4a). When C_{SW} is presented as a function of SST (Fig. 4b), the separate influences of the two defining factors, namely S_{\odot} and α , becomes apparent. In the winter hemisphere, C_{SW} decreases poleward since the available sunlight (S_{\odot}) decreases poleward. In the summer hemisphere, significant reflection of solar radiation occurs over the illuminated clouds in the midlatitude storm tracks and C_{SW} increases with decreasing SST. The correlation of the changes in cloud flux quantities with changes in SST is often taken to be a measure of cloud feedback (sections 8 and 9).

b. Interannual and decadal variability

Understanding and quantifying the reciprocal effects of clouds on the TOA radiation budget is fundamen-

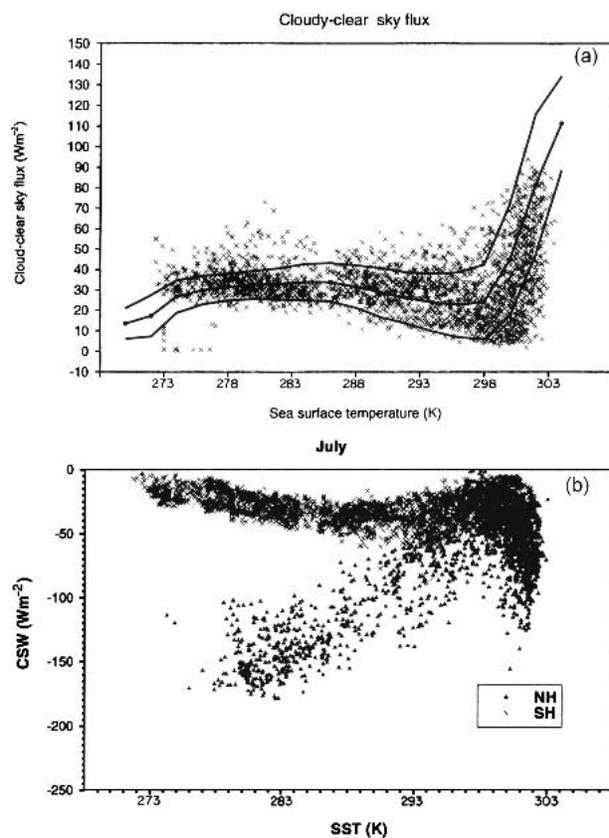


FIG. 4. (a) The cloud longwave forcing as a function of SST. The lines represent the comparison of a GCM model; its spread is defined as one std dev from the central line (Stephens et al. 1993). (b) The cloud shortwave forcing as a function of SST for Jul (Stephens et al. 1993).

tally important for a number of reasons. Notably, any change in cloud that alters one component without a compensating change in the other potentially induces significant changes to the net radiation balance. Most cloud feedback concepts are posed in this way although often without justification.

Different indices have been created over the years to demonstrate the nature and extent of the TOA cancellation (Ohring and Clapp 1980; Hartmann and Short 1980; Stephens and Greenwald 1991, and others). For illustration, consider the simple ratio

$$r = -\frac{C_{\text{SW}}}{C_{\text{LW}}}$$

as introduced by Cess et al. (2001). This ratio has the value $r = 1$ when shortwave and longwave effects precisely cancel each other out. Although this ratio typically varies between a value of 1 and 1.1 for much of the tropical atmosphere, Cess et al. (2001) reported on a significant interannual variability of this ratio (Fig. 5), which reflects changes to tropical circulation patterns (Allan et al. 2002).

Only recently has it become possible to observe the kinds of interannual variability in the ERB presented in Fig. 6. The availability of new satellite measurements during the past decade together with the sustained measurements from the wide field-of-view (WFOV) radiometer of ERBE represents an opportunity for piecing together TOA flux records extending over almost two decades. In constructing such a time series, Wielicki et al. (2002; Fig. 6) document a systematic change between 1985 and 2000 in TOA fluxes emerging from the tropical atmosphere. Their analysis points to changes in cloudiness over this period associated in an uncertain way with apparent changes to the meridional circulation of the atmosphere (e.g., Chen et al. 2002; Cess and Udelhofen 2003).

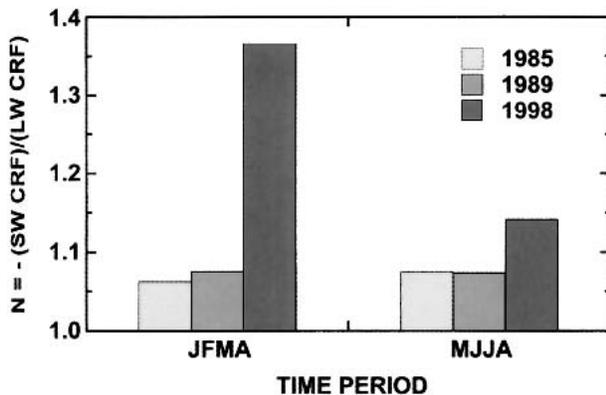


FIG. 5. The ratio of cloud radiative forcing presented for a small region of the equatorial Pacific that exhibits little interannual variability in SST. Cloud systems in this region during the Jan–Feb–Mar–Apr season of 1998 appear to have changed significantly from clouds of other years with ratio values $R \approx 1.35$ due in large part to a reduction in C_{LW} (Cess et al. 2001).

The variabilities observed by both Cess et al. (2001) and Wielicki et al. (2002) as well as those reported by Kuang et al. (1998) may well be relevant to the problem of cloud feedback since it appears to be a manifestation of the links between cloud, radiation, and the larger-scale circulation of the atmosphere proposed above as the framework to understand cloud–climate feedbacks (Fig. 3). These observed decadal and interannual variabilities require a more quantitative and deeper level of understanding of the connection between clouds and atmospheric circulation than presently exists.

c. Effects of clouds on the ATM and SFC radiation budgets

Most cloud feedback studies concern themselves with the effects of clouds on the TOA fluxes. Yet there are important potential feedbacks that are governed not by TOA flux effects but by the effects of clouds in heating and cooling the atmosphere. Understanding how clouds partition the absorption of radiation between the surface and atmosphere requires a global surface radiation budget climatology, in combination with TOA fluxes. A major obstacle in determining the radiation budgets of both the atmosphere and surface are the limitations of the input cloud properties and other information needed in radiative transfer calculations. Nevertheless, different global climatologies of the surface radiation budgets, based largely on satellite data, have been developed over the years (e.g., Zhang et al. 1995; Bishop et al. 1997; Gupta et al. 1999; Curry et al. 1999; L’Ecuyer and Stephens 2003; among others).

Given these climatologies, and despite their limitations, we can deduce that the net cooling effect of clouds, noted above in relation to the TOA budget, occurs broadly as a residual of cooling associated with effects of clouds on solar radiation at the surface and the warming of the atmosphere by the effects of clouds on longwave fluxes (e.g., Stephens et al. 1994; Rossow and Zhang 1995). Relative to clear skies, clouds heat the low-latitude atmosphere through a combination of increased IR absorption and emission at colder temperatures and cool the surface through reflection of solar radiation to space depleting the amount of solar radiation absorbed at the surface. The combination of these two effects produces the largely reciprocal cancellation observed at the TOA and discussed previously. By contrast, high-latitude clouds affect the radiation balance in a manner that is almost the reverse of the effect at low latitudes (e.g., Rossow and Zhang 1995; Stephens 2000). Thus clouds enhance the latitudinal gradient of column cooling and reinforce the meridional heating gradients responsible for forcing the mean meridional circulation of the atmosphere.

The extent to which cloud layers heat or cool the atmosphere (relative to clear skies) is also largely determined by the vertical location of clouds. High, cold clouds tend to warm the atmospheric column relative to

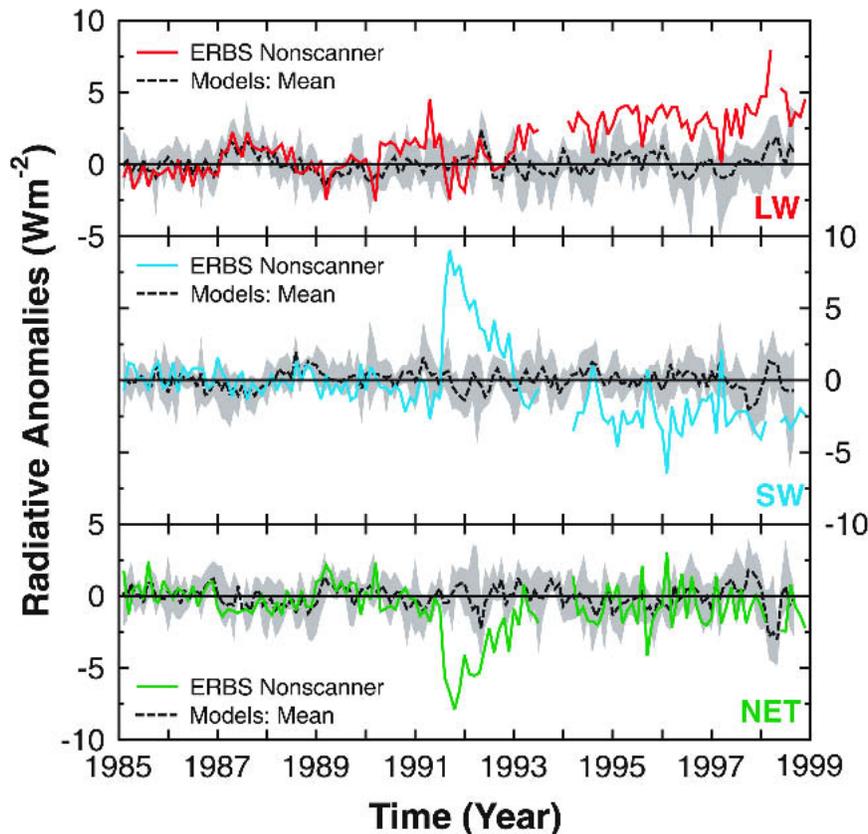


FIG. 6. Satellite record of tropical mean (20°S – 20°N) anomalies in broadband-emitted longwave fluxes, reflected solar fluxes, and net TOA fluxes. Results are representative of seven different instruments flown on six different satellites (Wielicki et al. 2002). The gray shading shows the same quantities derived from AMIP model integrations.

surrounding clear skies, particularly at low latitudes, whereas low clouds enhance the cooling of the atmosphere, particularly at high latitudes (e.g., Slingo and Slingo 1988). This introduces an additional and important dependence not evident in the above simple discussion of TOA fluxes.

Many of the effects of clouds on the surface and atmospheric radiation budgets are highlighted in Figs. 7a,b,c. Shown in this example is a composite view of the radiation and water budgets of the Madden–Julian oscillation (MJO) inferred from the Tropical Rainfall Measuring Mission (TRMM) observations. Details on how TRMM data are analyzed to produce the given parameters are described elsewhere (L’Ecuyer and Stephens 2003). Figure 7a presents the composite variations of cloud ice water path and precipitation expressed relative to the cycle of SST during the MJO. The maximum of this SST cycle corresponds to day 0 in the manner of Fasullo and Webster (1999). Figure 7b similarly presents the variation of reflected solar radiation, surface solar radiation, and outgoing longwave radiation (OLR) and Fig. 7c present the column atmospheric radiative heating and components of this col-

umn heating.² The notable features are the following: (i) The cycle of TOA reflected solar and OLR are largely reciprocal as already noted with elevated reflection and reduced OLR coinciding with the SST cooling phase, which happens to correspond to the period of deep convection, high precipitation, and extensive high cloud indicated by increased ice water path (IWP). (ii) The variation of the downwelling surface solar radiation largely mirrors the reflected TOA solar flux. Clouds produce little change to the column-mean-integrated solar absorption in the atmosphere. However, they do redistribute this absorbed radiation now being concentrated in cloud layers rather than spread throughout much of the lower half of the troposphere in clear skies (Stephens 1978). In fact, methods for estimating surface solar fluxes are based primarily on this

² Since the atmospheric column radiative flux divergence differs from the column radiative cooling by a factor that is more or less constant, the term “column radiative cooling” will be used interchangeably throughout to refer to either the column flux divergence/convergence (in W m^{-2}) or equivalent column radiative cooling/heating (in K day^{-1}).

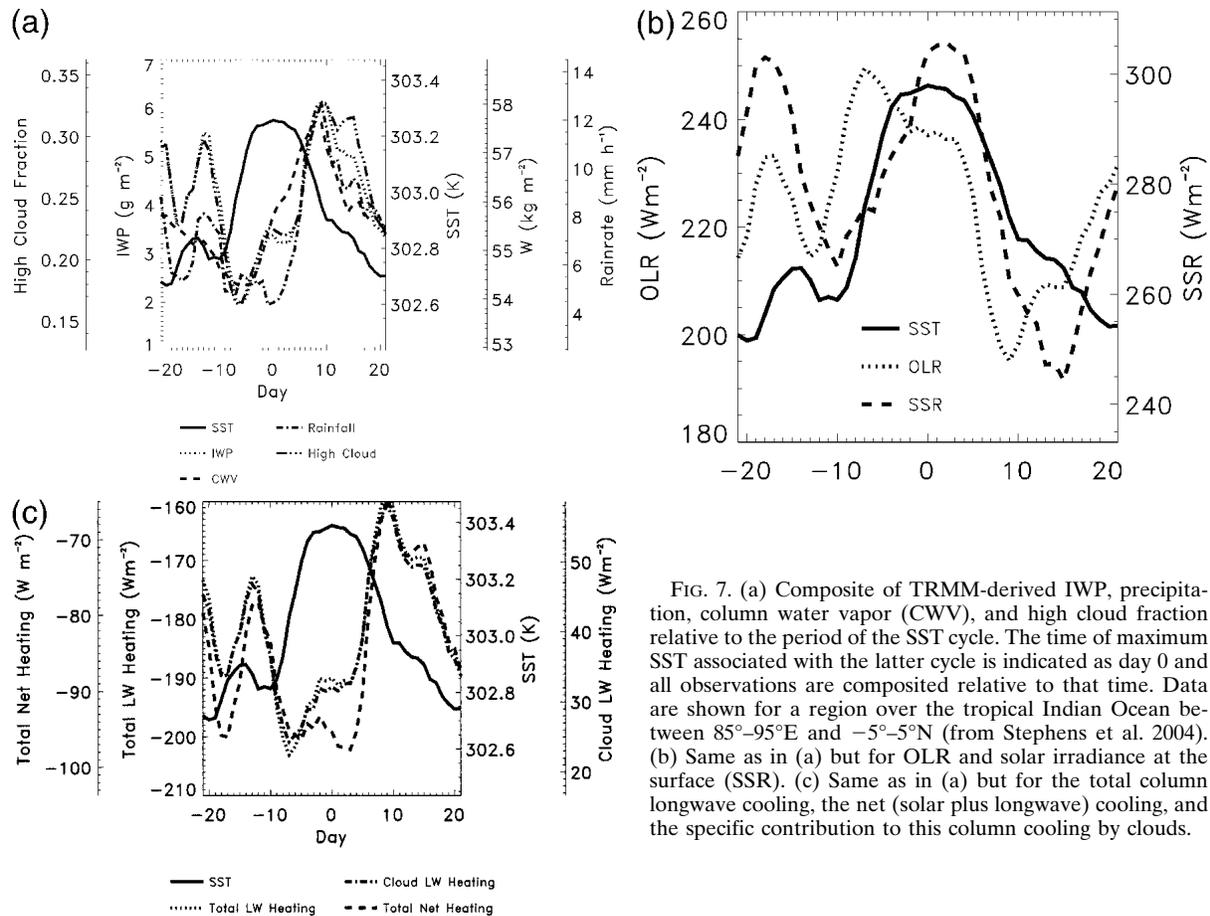


FIG. 7. (a) Composite of TRMM-derived IWP, precipitation, column water vapor (CWV), and high cloud fraction relative to the period of the SST cycle. The time of maximum SST associated with the latter cycle is indicated as day 0 and all observations are composited relative to that time. Data are shown for a region over the tropical Indian Ocean between 85°–95°E and –5°–5°N (from Stephens et al. 2004). (b) Same as in (a) but for OLR and solar irradiance at the surface (SSR). (c) Same as in (a) but for the total column longwave cooling, the net (solar plus longwave) cooling, and the specific contribution to this column cooling by clouds.

observation (e.g., Li and Leighton 1993). (iii) The atmospheric column cooling also varies in a manner closely correlated to this surface solar variation with minimum cooling (i.e., a maximum cloud heating relative to clear-sky cooling) occurring at the times of minimum surface solar flux. The fluctuation in column cooling also closely mimics the variation of IWP through the cycle.

d. ISCCP and the radiation budget

It is somewhat surprising that there are relatively few studies that seek to examine the relationships between the cloud flux quantities and other cloud properties in an effort to understand what radiation processes determine the observed effects of clouds on fluxes. Stephens and Greenwald (1991) is one example in which ERBE data are correlated with satellite microwave radiance data to explore the relation between both C_{LW} and C_{SW} and the cloud liquid water path, a relationship central to the cloud optical depth feedback described later.

The majority of the research on this topic has mainly focused on relations between TOA cloud fluxes and ISCCP cloud parameters. At this point it is helpful to consider the key outputs of the ISCCP. The principal

cloud properties derived by ISCCP are the cloud optical depth (τ), cloud-top pressure (CTP), and cloud amount. A compact way of presenting these particular ISCCP products is in the form of the 2D histograms shown in Fig. 8a, which represent cloud amount information grouped into nine basic categories loosely identified by cloud types (Hahn et al. 2001). More specifically, these histograms are constructed from a total of six different τ ranges and seven different CTP levels thereby creating 42 different τ –CTP specific categories. As discussed later, this organizational model is becoming widely used in studies that explore the relationships between clouds and other parameters of the climate system. This concept too has recently been extended to climate model data analysis (e.g., Webb et al. 2001) and community ISCCP simulators exist for this purpose (see online at <http://gcss-dime.giss.nasa.gov/simulator.html>).

Clearly the above-mentioned energy balance studies of Zhang et al. (1995) and the follow-up study of Chen et al. (2000) are important examples of research that connect ISCCP cloud properties to TOA-, SFC-, and ATM-derived radiative fluxes and thus provide a basis for understanding these connections. One of the earliest studies of this type is that of Okhert-Bell and Hart-

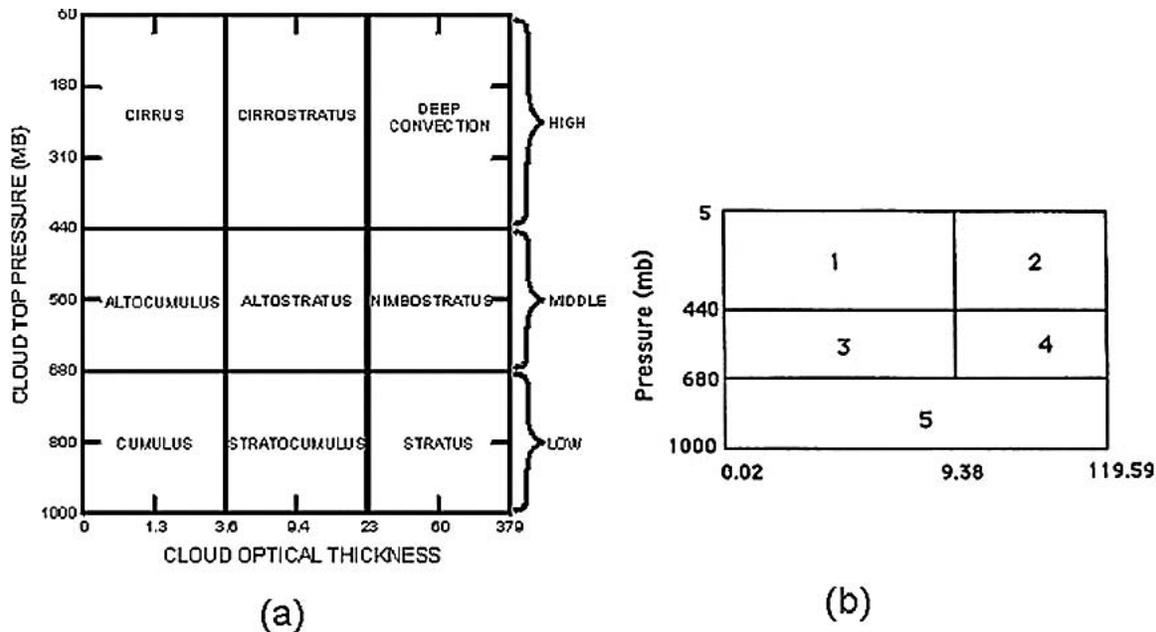


FIG. 8. (a) The radiometric classification of cloudy pixels in terms of optical depth and cloud-top pressure according to ISCCP. (b) The reduced set of five cloud types of OBH.

man (1992, hereafter OBH). They introduced a simple method for correlating ISCCP and ERBE data to determine what types of clouds contribute most to the observed C_{LW} and C_{SW} . They introduced these cloud fluxes as

$$C_{\text{SW,LW}} = \sum_{i=1}^5 \Delta F_i N_i, \quad (5)$$

where ΔF_i is the change in the relevant TOA flux associated with overcast cloud of type i and N_i is the fractional cloud coverage by the i th category. In this approach OBH consider a reduced set of cloud categories (five in all as shown in Fig. 8b) and their regression analysis determines ΔF_i given $C_{\text{SW,LW}}$ from ERBE and N_i from ISCCP. Table 1 summarizes the results of the OBH analysis presenting the global-mean values of ΔF_i . The dominance of the high clouds (types 1 and 2 in Fig. 8b) to the OLR at least in low latitudes emerges from the analysis, as does the importance of the thicker high

clouds in the Tropics and low clouds in mid- to high latitudes on shortwave fluxes. Net fluxes are influenced most by low clouds especially through their effect on solar radiation in the summer hemisphere. Chen et al. (2000) extend the analysis of OBH to the ATM and surface budgets.

4. The radiative-convective equilibrium paradigm

Early estimates of the surface temperature warming induced by increasing in CO_2 were based on an assumption of simple radiative equilibrium (e.g., Moller 1963). These early estimates, however, were implausibly large because of the overly simple nature of this assumption (e.g., Held and Soden 2000). To first order, the atmosphere exists in a state of quasi balance between radiative cooling and the convective processes that give rise to latent and sensible heating.

Although the processes that establish this quasi-

TABLE 1. The contributions to the LW, SW, and net cloud forcings by the five different cloud classes identified by OBH and Fig. 7a. The cloud amount (N) for each type is also given (from Hartmann et al. 1992).

	Type 1 high, thin		Type 2 high, thick		Type 3 mid, thin		Type 4 mid, thick		Type 5 low		Sum Avg
	JJA	DJF	JJA	DJF	JJA	DJF	JJA	DJF	JJA	DJF	
N	10.2	10.0	8.5	8.8	10.7	10.7	6.5	8.2	27.2	25.9	63.3
OLR	6.5	6.3	8.4	8.8	4.8	4.9	2.4	2.4	3.5	3.5	25.8
Albedo	1.2	1.1	4.1	4.2	1.1	1.0	2.7	3.0	5.8	5.6	14.9
Net	2.4	2.3	-6.4	-7.5	1.4	0.8	-6.6	-8.5	-15.1	-18.2	-27.6

radiative–convective equilibrium (RCE) are complex, simple, approximate ways of circumventing this complexity were introduced in a series of papers by Manabe and colleagues (e.g., Manabe and Strickler 1964; Manabe and Wetherald 1967). Their method added a fixed lapse rate constraint to the 1D radiative equilibrium model used in the earlier study of Manabe and Moller (1961) as a way of approximating the effects of convection. This model was then able to realistically predict the position of the tropopause below which a prescribed lapse rate is maintained.

This simple model served as a valuable point of reference for more than 20 years and much was learned from studies that employed these models.

a. RCE and cloud feedback

Manabe and Strickler (1964) were the first to point out via their simple RCE model that high cirrus clouds heat the surface by an amount affected by their height and emissivity whereas low clouds cool the surface (Fig. 9). They argued that this heating occurs when the emissivity exceeds about 0.5 (i.e., thicker cirrus). They refer to this as the critical blackness but it was later shown that this idea is based on an unrealistic assumption that the albedo remains fixed as the emissivity increases. The fact that the albedo and emission from clouds are

somehow related producing a complicated balance at the TOA was not understood at that time. When treated in a physically consistent manner, linked via cloud water and ice paths and thus optical depth as introduced first by Stephens and Webster (1981), then it is not the thicker cirrus that produces a surface heating but rather the thin cirrus (Fig. 9b).

The motivation for the use of simple RCE models in these kinds of experiments had less to do with defining actual cloud feedback and climate sensitivity per se but more to do with demonstrating the potential relevance of cloud–radiation interactions to the climate system. As noted previously, the decade of the 1970s was a period of some debate about the sensitivity of climate to cloud and these equilibrium studies, followed later by GCM studies, eventually elevated the general problem of cloud feedback to a level of high importance. Cloud feedback concepts were also specifically pursued using RCE models. Paltridge (1980), Wang et al. (1981), Charlack (1982), Somerville and Remer (1984), Somerville and Iacobellis (1986), and Stephens et al. (1990) are all examples of cloud feedback ideas that involve cloud optical depth and equivalent water and ice path information. A more systematic analysis of the water path–optical depth feedback is provided below.

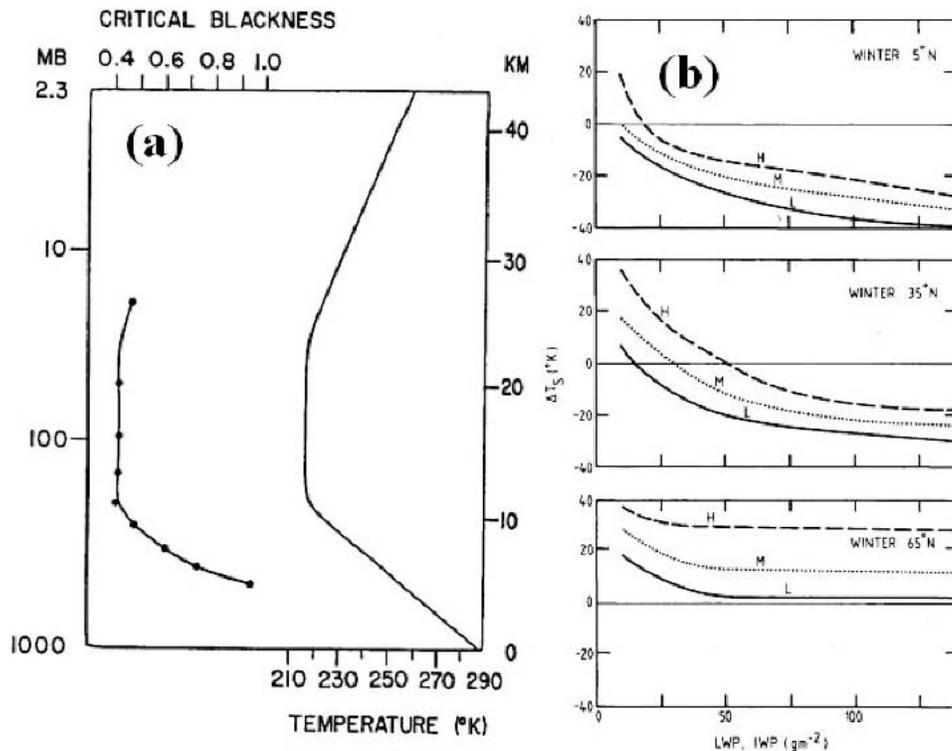


FIG. 9. (a) The critical-blackness cloud experiment of Manabe and Strickler (1964). The profile to the left is the profile of emissivity above which clouds warm the surface relative to the given clear-sky temperature profile (right). (b) The change in equilibrium surface temperature as a function of cloud LWP and IWP for low (L), middle (M) and high (H) clouds (Stephens and Webster 1981).

b. RCE and cloud-resolving models

The early RCE studies were meant to represent, loosely, a quasi-global-mean state. These simple models have generally fallen out of favor being replaced by more complex GCM climate models although many of the sensitivities derived from RCE models were broadly replicated in GCMs (Roeckner 1988; LeTret and Li 1991; and others). More recently, however, the RCE paradigm has been revisited in a series of equilibrium experiments conducted using cloud-resolving models (CRMs). The focus of these studies was directed toward convection in the tropical atmosphere.

The use of CRMs in these experiments offers a more self-consistent treatment of convection and related cloud–radiation processes than is possible with the simple RCE models although it might be argued that the resolution of the CRMs used, typically $\Delta x \sim 2\text{--}4$ km, only marginally resolves convection. Equilibrium integrations reported are also mainly for CRMs set on a 2D domain representing a vertical slice through the atmosphere (Held et al. 1993; Sui et al. 1994; Grabowski 2001) whereas the 3D model experiments performed by Tompkins and Craig (1998) were, necessarily, set on a more limited domain.

Different model domain sizes together with varying levels of sophistication of both cloud microphysics and the coupling of radiation–cloud processes make it difficult to draw general conclusions about the equilibrium reached and the relation of this equilibrium to the real atmosphere. However certain general features appear to be robust, including the following: (i) Feedbacks are set up between the distribution of water vapor and convection that result in a mutual organization of convection. (ii) These feedbacks are organized to a large degree by secondary circulations driven by radiation differences between convective regions and clear-sky regions of subsidence much in the manner hypothesized by Gray and Jacobsen (1977). (iii) A characteristic time scale of feedbacks seems to emerge established by the radiatively driven subsidence of the clear skies.

To date, CRM–RCE experiments have been constructed as open systems of fixed SST (e.g., Tompkins 2001; Grabowski et al. 2000; and also Fig. 10a and related discussion). Feedbacks discussed in these studies refer to interactions between cloud and convection processes and the large-scale environment in which these processes take place. The dependence of these cloud and convection processes on temperature and the specific coupling between these processes to SST, more in the mode of a closed climate feedback system (Fig. 10b), has yet to be studied.

c. RCE and the earth's hydrological cycle

RCE can also be defined in terms of the atmospheric energy budget. This budget, to first order, occurs as a balance between the radiative cooling of the atmosphere and latent heating associated with precipitation.

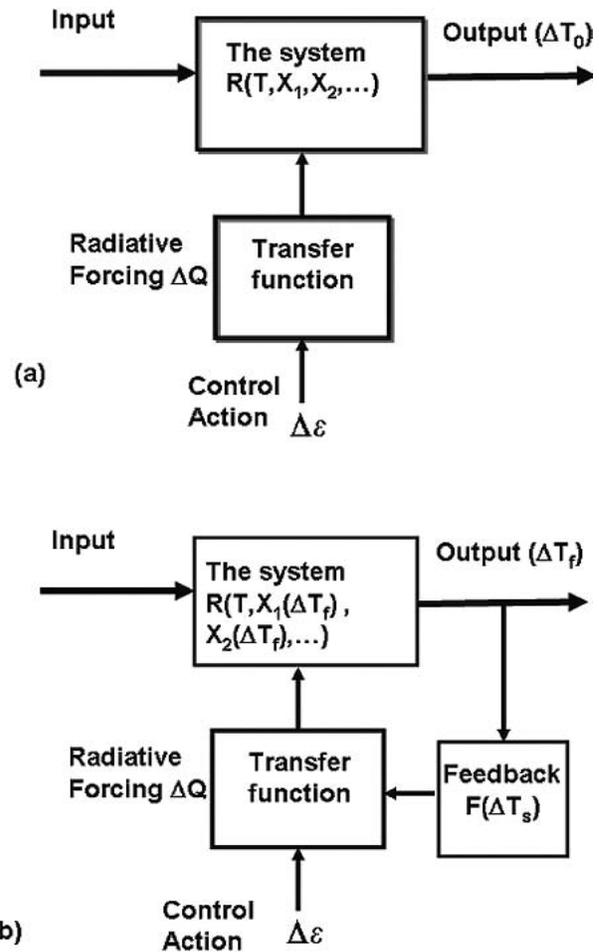


FIG. 10. (a) The main elements of an open (energy balance) climate system, where X is independent of output ΔT_0 . The climate forcing involves the transfer function mapping from the control action to a change in radiative budget. (b) The control action modulates the output ΔT_f and feedback is established by a loop connecting the output to input.

Thus RCE implies that the radiation balance of the atmosphere and the planetary hydrological cycle are connected. With the Renno et al. (1994) study aside, the early, simple RCE models include effects of convection only approximately via the convective adjustment approximation with no relation to moist convection and precipitation. In principle, the link between radiative cooling of the atmosphere, convection, and precipitation emerges in a more self-consistent and realistic manner in the more recent RCE–CRM models.

Given this approximate balance, it might then be argued that changes to the radiative cooling of the atmosphere fundamentally establishes changes to the hydrological cycle, at least on some large scale. For example, Gray and Jacobsen (1977), Fingerhut (1978), and McBride and Gray (1980) argue that the diurnal cycle of precipitation is a product of the day–night differences of atmospheric radiative cooling. Mapes (2001)

proposes that clear-sky radiative cooling through the subsidence it produces governs the profile of water vapor especially in the upper troposphere. Hartman and Larson (2002) propose that the level of detrainment of tropical convective clouds occurs at levels where the clear-sky radiative cooling decreases rapidly near 200 hPa, the latter being fundamentally controlled by the nature of the vertical distribution of upper-tropospheric water vapor. These authors further suggest that the emission temperature of these anvil clouds remains essentially fixed being independent of SST and, by implication, any effects of climate change. This argument, however, hinges on the assumption that the relative humidity in the clear-sky portions of the atmosphere where subsidence occurs is fixed and independent of SST.

If one accepts the simple hypothesis that the hydrological cycle adjusts to changes in the atmospheric radiative cooling (e.g., Mitchell et al. 1987; Stephens et al. 1994; Sugi et al. 2002), then we have a basis for interpreting how the hydrological cycle might change under global warming. As noted by Stephens et al. (1994) and evident in Fig. 7 and later in Fig. 12, both column water vapor and clouds are principal factors that influence the gross radiative budget of the atmosphere. Consequently, changes to clouds and water vapor that induce a change to the column atmospheric cooling will, in turn, produce compensating changes to the hydrological cycle. It is through this connection that cloud feedbacks become relevant not only to the problem of surface warming (Fig. 1) but also to the problem of the response of the planet's hydrological cycle to global warming (refer to Fig. 15 and related discussion).

5. Cloud feedbacks and the regulation of tropical SSTs

A number of studies suggest that tropical SSTs come under the influence of a runaway water vapor green-

house effect and that a negative feedback must operate to limit the climatological SSTs to about 30°C. This point was notably raised in the observational study of Ramanathan and Collins (1991) who referred to this runaway effect as the "supergreenhouse" effect and the regulation of SSTs as the thermostat hypothesis. The idea of a runaway greenhouse effect in the absence of a regulatory negative feedback is also supported by simple energy balance arguments (e.g., Pierrehumbert 1995; Kelly et al. 1999; among others).

Over the years, a number of different hypotheses about the nature of this negative feedback have been proposed. These hypotheses are broadly grouped into three categories:

- 1) The concept for the negative feedback in this first category of study centers around the mechanism associated with evaporation established by large-scale winds (e.g., Newell 1979; Priestly 1964) and pursued further by Bates (1999). Using a simple energy balance model, Bates (1999) illustrated how a feedback induced by the coupling of meridional momentum transport, low-level winds, and evaporation in principle is sufficient to maintain tropical SSTs near the values observed.
- 2) The second category deals with feedbacks that combine the radiation balance and large-scale dynamics. Pierrehumbert (1995), Miller (1997), and Larson et al. (1999) all argue that large-scale dynamics produces a communication between the atmosphere above the warmest waters and deepest convection and the atmosphere above the cooler waters as part of the Walker circulation. Pierrehumbert (1995) proposes that heat is transported from the convective region over warm SSTs to the regions of subsidence over the cooler waters where heat escapes by elevated levels of emission to space. Miller (1997) and Larson et al. (1999) suggest that a negative feed-

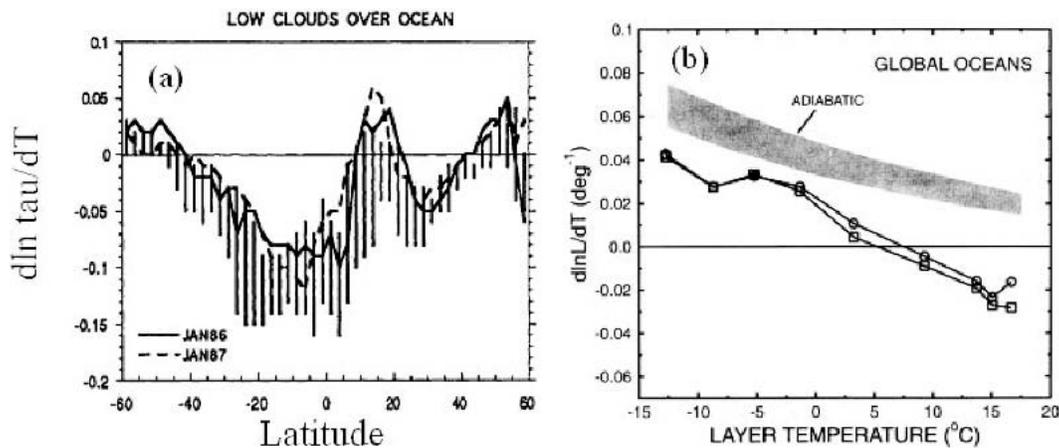


FIG. 11. (a) The local sensitivity of cloud optical depth derived from ISCCP with cloud temperature (Tselioudis and Rossow 1994). (b) The local sensitivity of cloud liquid water path derived from SSM/I microwave radiance data with cloud temperature (Greenwald et al. 1995).

back arises via the communication between these two regions such that increases in convection in the warm pool region indirectly lead to increased areal coverage of these low clouds in the cold pool region.

- 3) The third category of feedback, about which much has been written in recent times, proposes a cloud–radiation feedback involving the relationships between SST, deep convection, and detrained anvil cirrus and solar radiation. For example, Graham and Barnett (1987), Ramanathan and Collins (1991), and Chou and Neelin (1999) all propose that reduced insolation in regions of extended cirrus cloud cover associated with convection produces the fundamental SST-stabilizing mechanism. Ramanathan and Collins (1991) attempt to examine this feedback by contrasting observations between an El Niño and non–El Niño year arguing that this contrast provides a natural experiment of climate change. Their analysis focuses on an observed correlation between changes in C_{sw} and SST between the April months of these two years. Fu et al. (1992) point out that the correlations are not robust, a claim later supported by other model and observational studies that suggested that C_{sw} is better correlated with large-scale divergence than with SST (Hartmann and Michelson 2002; Lau et al. 1996).

The debate over the role of evaporation and other large-scale effects versus more localized cloud–radiation processes continued with Wallace (1992), Waliser and Graham (1993), and Waliser (1996). Stephens et al. (1994) demonstrated how regions of supergreenhouse effects over high SSTs are also regions of significantly increased radiative cooling of the atmosphere in the absence of clouds implying that this increased cooling supports convection via atmospheric destabilization effects (e.g., Hartmann and Larson 2002; Stephens et al. 2004).

Recently, Lindzen et al. (2001) claim that increased convection associated with increased SST eventually leads to a decrease in cirrus, in direct contrast to the underlying hypotheses of the feedback studies noted under (iii) above. Lindzen et al. (2001) invoke the results of earlier RCE studies that suggest increased thin cirrus warm the surface temperatures (e.g., Fig. 8b), thus implying that SSTs must cool when this thin cirrus is reduced. There are a number of inconsistencies in this “adaptive iris” hypothesis as noted by DelGenio and Kovari (2002), Hartman and Michelson (2002), and Lin et al. (2002). However, the study raises an important point about the relevance of precipitation efficiency and how this efficiency might change with SST (and, by implication, climate change). Precipitation efficiency in this context loosely refers to the proportion of cloudiness (cirrus) to the proportion of precipitation (convection). As mentioned above and shown later, changes to the global precipitation are governed by changes to the atmospheric radiative cooling, the latter

being significantly influenced by clouds. It follows then that the precipitation efficiency is grossly controlled by changes in the vertical distribution of clouds through the effects of this distribution on the radiation cooling of the atmosphere.

In one way or another inconsistencies can be identified with all SST regulation concepts described above either as a result of overly simplistic theories that overlook or ignore processes without justification, or through the lack of conclusive observations that might confirm the essential assumptions of these theories. DelGenio and Kovari (2002), Lau et al. (1996), Bony et al. (1997) and Hartmann and Michelson (2002) all note that the local SST dependence of single parameters considered proxies for convection (e.g., cirrus amount, precipitation, OLR) are affected by a host of other parameters and thus, by implication, other processes. Simple theories generally assume that only one process of the coupled system dominates over all others and thus a priori assert a cause and effect. Unfortunately, it is practically impossible to verify the simplifying assumptions of the hypotheses in part because we cannot isolate those processes in the real world and in part due to the ambiguous nature of the proxy data used to examine processes. An example of such ambiguity is provided in comparison of the studies of Chou and Neelin and Lindzen et al. Both studies use essentially the same raw satellite data yet reach opposite conclusions about the relation between convection, cirrus, and SSTs.

6. A systems perspective of cloud feedback

It was remarked in section 2 how control theory has been used as a guiding framework for analyzing climate feedbacks (e.g., Schlessinger and Mitchell 1987; Arking 1991; Curry and Webster 1999), and most notably for analyzing feedbacks in GCMs. For this reason and to provide more substance to the discussion of section 2, a brief review of the system approach is presented. It is also useful to contemplate the simple RCE and SST feedback studies described above in this systems framework and draw from it lessons learned about the shortcomings of such studies.

A control system here is defined as an arrangement of connected physical components that act as an entire self-regulating unit. Four components define a control system—the input, the output, which is stimulated by the input, and the “system” that defines the relation between output and input (Fig. 10a). We might think of the system as the entire global climate system composed of subsystems with connecting inputs and outputs. The input is the solar energy received from the sun and this input can vary on many time scales (diurnal, seasonal, and longer). It is important to understand what feedbacks operate and how they operate as the input conditions change. The output of the system may also be expressed in a number of ways, but usually this

is taken to be the global-mean surface temperature T_s . The fourth component is the control action, $\Delta\epsilon$, which is responsible for activating the system to modulate the output now expressed as a change ΔT_s . Thus, $\Delta\epsilon$ represents the external changes to the system, such as an increase in atmospheric CO_2 concentration.

Figure 10 illustrates two types of control systems of heuristic relevance to climate feedbacks. The first (Fig. 10a) is an open system in which the control action is independent of output and the sensitivity λ_o , or alternatively the system gain λ_o^{-1} , is that which occurs *without* feedbacks. Note that the control action, $\Delta\epsilon$, connects to the system via a transfer function representing the procedures for converting $\Delta\epsilon$ into a climate radiative forcing ΔQ . There is extensive literature on the topic of climate radiative forcing. The second type of system (Fig. 10b) is a closed or feedback system wherein the the action triggers a response that modulates the radiative forcing. These feedbacks might operate in series where the output(s) of one feedback drives the input(s) of another feedback, or they might operate in parallel where all feedbacks are thought to operate independently of one another, or they might operate as some combination of both (e.g., Aires and Rossow 2003).

The sensitivity of a feedback system is λ_f and the feedbacks that define it can be quantified in a simple way by a side-by-side comparison of the closed system and its equivalent open system according to

$$\frac{\lambda_o}{\lambda_f} = 1 - f, \quad (6)$$

where the meaning of f becomes apparent below.

The use of this simple, heuristic system framework raises a number of questions when applied to the climate system (e.g., section 2 above and also Aires and Rossow 2003).

- 1) What is the system, its component processes, and its “control action”? As we have noted, hypothesized climate feedback systems such as those already described above are overly simple. The extent to which the real climate system resembles such simple systems (e.g., top part of Fig. 2) is questionable at best and requires, at the very least, a higher level of justification and some level of verification than is usually given. Verification is most problematic given that it is not obvious how we use observations to confirm the identity of the system. The identity of the climate system is itself the major source of confusion and uncertainty in feedback analysis.
- 2) What is the system output? Obviously feedbacks are only meaningful when defined with respect to a given output. As mentioned, the output assumed in feedback diagnostic studies is almost always taken to be global-mean surface temperature (e.g., Allen and Ingram 2002). It is not obvious that this is the most meaningful output and other metrics of the

climate system could certainly be considered. There is also no reason to suppose that only one output defines the system. Either way, different outputs or combinations of outputs naturally define a different system and different forms of feedback.

- 3) How do we observe or otherwise quantify open and closed system responses in parallel? Quantifying feedback requires knowledge about the (open) system free of the feedbacks under study, and determining this sensitivity is unfortunately more complex than is typically considered in most analysis. It is unlikely that the open-loop sensitivity can be derived from observations alone since we cannot generally observe the climate system with selected feedbacks turned off. Similarly, the use of models alone in feedback diagnostic studies without ties to observations has little value. Diagnostic methods that tie one to the other are essential if progress is to occur.

While there are a number of questions that continue to be raised about the applicability of the particular application of the systems approach to the study of climate feedbacks, the remainder of this review discusses the general nature of the approach as it has most commonly been used to study cloud feedbacks to date. This serves a useful purpose as it provides a reference for the discussion of section 2 and for the reference for contemplating the GCM analyzes described in section 8.

a. A simple open-loop system

To place the above discussion into the context of most climate model analysis, suppose the climate system is a collection of processes that forms the radiation budget at the top of the atmosphere R_{TOA} and suppose this budget can be defined uniquely by time- and space-averaged quantities:

$$R_{\text{TOA}} = (\epsilon, T, X_1, X_2, \dots) = 0, \quad (7)$$

where ϵ is a parameter that establishes the control action and T is the output. The variables X_1, X_2, \dots are processes that have no dependence on the output but necessarily define the system. The interpretation of this temperature output, at this point, is left vague. Here R_{TOA} defines the system connecting (solar) input to output (emitted radiation related to temperature). Although the direct relationship between R_{TOA} and T is sometimes inappropriately referred to as the temperature feedback (e.g., Colman et al. 1997), it is not a feedback in the context of a control system and is common to both open and closed systems. The relationship between R_{TOA} and T governs the open-loop gain λ_o^{-1} [Eq. (11)].

To fix ideas, consider a climate forcing established by the control action $\Delta\epsilon$, then

$$\Delta R_{\text{TOA}} = \frac{\partial R_{\text{TOA}}}{\partial \epsilon} \Delta\epsilon + \frac{\partial R_{\text{TOA}}}{\partial T} \Delta T. \quad (8)$$

Under the condition of equilibrium, $\Delta R_{\text{TOA}} = 0$, and

$$\Delta T_{\text{eq},o} = - \left(\frac{\partial R_{\text{TOA}}}{\partial T} \right)^{-1} \frac{\partial R_{\text{TOA}}}{\partial \epsilon} \Delta \epsilon \quad (9)$$

or

$$\Delta T_{\text{eq},o} = \Delta Q_o / \lambda_o, \quad (10)$$

where

$$\lambda_o = - \left(\frac{\partial R_{\text{TOA}}}{\partial T} \right) \quad (11)$$

and

$$\Delta Q_o = \frac{\partial R_{\text{TOA}}}{\partial \epsilon} \Delta \epsilon \quad (12)$$

is the climate “forcing” representing in the usual way an instantaneous change to the TOA radiation budget due to the control action. Here $\Delta T_{\text{eq},o}$ is the open system response. It is to be stressed that (1), (10), and the equivalent sensitivity of a feedback system (18) are valid only for stationary systems in equilibrium.

In contemplating the sensitivity λ_o , consider the simple system

$$R_{\text{TOA}} = \frac{Q_o}{4} (1 - \bar{\alpha}) - \sigma T_p^4 \quad (13)$$

for which the output in this example is the planetary temperature T_p . For this system $\lambda_o = 4\sigma T_p^3$ and $\lambda_o = 3.68 \text{ W m}^{-2} \text{ K}^{-1}$ for $T_p = 253 \text{ K}$. The relevance of this simple analysis to more complex systems such as the real climate system or even of climate simulated by GCMs is unclear being a gross oversimplification of these systems. To emphasize this point, consider a *different* system defined by the output T_s rather than T_p , then

$$\lambda_o = 4\sigma T_p^3 \frac{dT_p}{dT_s}, \quad (14)$$

where the functional dependence $T_p(T_s)$ represents the sequence of atmosphere processes and radiative transfer that establishes the connection between surface temperature T_s and T_p . The relationship between these temperatures, in reality, is complex and not easily determined for a climate system like that of the real earth. Stephens and Webster (1984) argue that the relationship between T_p and T_s , due to the effects of clouds on the radiation balance, may not even be unique.

An alternative approach invokes the relationship for OLR (e.g., Budyko 1969):

$$F_{\text{LW}} = A + BT_s, \quad (15)$$

where it follows from (11) and (13) that $\lambda_o = B$. The coefficients of the Budyko relationship are typically derived from spatiotemporal OLR and SST data, in which

case $B \sim 2 \text{ W m}^{-2} \text{ K}^{-1}$ (e.g., North et al. 1981). This obviously does not represent the actual sensitivity of the climate system with feedbacks turned off.

Yet a third approach to estimate the sensitivity of the open system free of cloud feedbacks is implied in the analysis of Cess et al. (1990) and the subsequent study of Arking (1991). This approach implicitly equates λ_o to the sensitivity derived using the clear-sky portions of the R_{TOA} in (11). Again there is no a priori reason to expect that this represents the system without cloud feedbacks given that clear-sky water vapor and other properties in the real world are influenced by the properties of the adjacent cloud skies through the circulations that connect one to the other.

The above are selected examples of methods used in feedback analysis to estimate λ_o . These approaches are attractive for their simplicity and for the fact that they seek to make use of observations. As we will see, the problems in specifying the open system sensitivity have greatly distorted the quantitative analysis of feedbacks. Quantifying λ_o strictly requires observations of the equivalent system without feedbacks, which obviously requires a clear identification of the system itself and some ability to isolate processes within it. Since we cannot generally observe the real climate system with feedbacks turned off, any use of observations for this purpose requires assumptions that are generally hard to justify a priori.

b. A simple example of a closed feedback system

Figures 1 and 10 conceptually point to how feedbacks alter the fundamental relationship between input and output and thus fundamentally determine the system sensitivity λ_f . It is simple to illustrate how feedbacks might alter the global-mean climate sensitivity by considering a *different* system

$$R_{\text{TOA}}[\epsilon, T, X_j(T), j = 1 \dots n] = 0, \quad (16)$$

where X_j are the processes that, in part, constitute the system. A process is defined here in terms of a dependent parameter X_j and an independent output variable T . Feedbacks are then established by the processes $X_j(T)$ that provide the return portion of the feedback loop. For example, consider a system of n feedbacks each operating *independently* of the other. In retracing the steps from (7) to (10) we obtain

$$\Delta R_{\text{TOA}} = \frac{\partial R_{\text{TOA}}}{\partial \epsilon} \Delta \epsilon + \left(\frac{\partial R_{\text{TOA}}}{\partial T} + \sum_{j=1}^n \frac{\partial R_{\text{TOA}}}{\partial X_j} \frac{dX_j}{dT} \right) \Delta T, \quad (17)$$

and with the explicit assumption of equilibrium

$$\Delta T_{\text{eq},f} = \Delta Q_o / \lambda_f, \quad (18)$$

where ΔQ_o is the forcing as defined above, $\Delta T_{\text{eq},f}$ is the response of the feedback system, and

$$\lambda_f = - \left(\frac{\partial R_{\text{TOA}}}{\partial T} + \sum_{j=1}^n \frac{\partial R_{\text{TOA}}}{\partial X_j} \frac{dX_j}{dT} \right). \quad (19)$$

Equation (6) follows from the ratio of (18) and (10):

$$\frac{\Delta T_{\text{eq},f}}{\Delta T_{\text{eq},o}} = \frac{\lambda_o}{\lambda_f} = 1 - f,$$

where from the definition

$$f = \frac{1}{\lambda_o} \sum_{j=1}^n \frac{\partial R_{\text{TOA}}}{\partial X_j} \frac{dX_j}{dT} = f_1 + f_2 + f_3 + \dots \quad (20)$$

f now emerges as a feedback parameter such that

$$f_j = \frac{1}{\lambda_o} \frac{\partial R_{\text{TOA}}}{\partial X_j} \frac{dX_j}{dT} \quad (21)$$

is a measure of the feedback arising from process $X_j(T)$. Note that λ_f contains both the sensitivity of the open system (i.e., $\partial R_{\text{TOA}}/\partial T$) and the accumulated effects of feedbacks expressed by f . As such λ_f does not quantify the feedbacks directly. For an open-loop system, $f = 0$ because $dX_j/dT = 0$ and $\lambda_f/\lambda_o = 1$.

A strategy typically employed to estimate f from model simulations is to determine each individual feedback contribution f_i and then sum these according to (21) (e.g., Arking 1991; Lindzen et al. 2001; Tsushima and Manabe 2001; Paltridge 1991; and many others). To do so requires quantitative estimates of the individual sensitivities factors $\partial R_{\text{TOA}}/\partial X_j$. Two model-based methods have been introduced for this purpose. One uses a simple one- or two-dimensional model (e.g., Hansen et al. 1984) and the second, introduced by Wetherald and Manabe (1988), uses 3D GCM-derived fields as input into offline radiation calculations. Soden et al. (2004) refer to the latter method as the partial radiative perturbation method, which typically considers the TOA budget in the form

$$R_{\text{TOA}} = R[X_1(T_s), X_2(T_s), \dots, X_n(T_s); T_s],$$

where $X_{1,\dots,n}(T_s)$ represent the physical processes that connect a given variable such as cloud type and amount, water content, lapse rate, water vapor, etc., to surface temperature (e.g., Zhang et al. 1994; Colman et al. 1997, 2001). The approach is then to replace an identified set of parameters X_j one at a time with the new set of GCM parameters $X_j + \delta X_j$ derived from a forced $2X\text{CO}_2$ or SST perturbation experiments. This produces a change in the TOA radiation budget at any model grid point as

$$\begin{aligned} \delta R_j &= R[X_1(T_s), \dots, X_j(T_s) + \delta X_j, \dots, X_n(T_s); T_s] \\ &\quad - R[X_1(T_s), X_2(T_s), \dots, X_j, \dots, X_n(T_s); T_s], \end{aligned}$$

and a total perturbation as

$$\sum_{j=1}^n \delta R_j = \sum_{j=1}^n \left(\frac{\partial R_{\text{TOA}}}{\partial X_j} \frac{dX_j}{dT} \right) \delta T_s. \quad (22)$$

For SST perturbation experiments like those described in section 8, δT_s is essentially prescribed and given the evaluation of the lhs of (22) by the Wetherald and Manabe (1988) method of substitution just described, a simple regression of one against the other provides an estimate of the factor

$$\sum_{j=1}^n \left(\frac{\partial R_{\text{TOA}}}{\partial X_j} \frac{dX_j}{dT} \right)$$

and thus a measure of feedback. Although these perturbations are derived at each grid point locally, using near-daily fields, they are typically analyzed in the terms of global-time-mean quantities in an attempt to estimate the global strengths of feedbacks.

Apart from the main problem with these kinds of analyses, that of system identification, there are other problems in the way the analyses are usually implemented. In applying the methods to estimate the individual feedbacks strengths f_i , it is usually assumed that the feedbacks are both independent of each other and are linear in nature. Wetherald and Manabe (1988) found that the use of time averages distorted the analyzed feedbacks due to nonlinearities inherent in cloud-radiation processes. Taylor and Ghan (1992) noted that temporal and spatial averages of cloud water also distorts the interpretation of feedback. Ingram et al. (1989) point out that care must be taken when evaluating albedo feedbacks in terms of area and time averages. Colman et al. (1997) extended the analysis procedures described above to include the second-order differentials of the form

$$\delta X_i = a_i \delta T_s + b_i (\delta T_s)^2 + e_i,$$

and estimated the magnitudes of these nonlinear terms for the $\pm 2\text{-K}$ SST experiments. They find that the largest nonlinear perturbations of the radiation budget response to parameter perturbations were those associated with lapse rate and high cloud and proposed that all other feedbacks are characteristically linear.

The assumption that feedbacks operate independently of one another is also usually done out of convenience. Zhang et al. (1994) suggested that cloud feedbacks alter the strength of the lapse rate feedback and Colman et al. (2001) extend the analysis approach of Wetherald and Manabe (1988) to include contributions by coupled feedback processes as a residual term in the estimated sensitivity. This analysis pointed to a distinct coupling between water vapor and cloud feedbacks as expected.

c. The cloud optical depth feedback example

As suggested in the previous section, a number of studies indicate that cloud feedbacks couple to other

feedbacks in the climate system making analysis more complicated than merely determining the individual feedback parameter f_i . The principal process connecting cloud feedbacks to others, and water vapor feedbacks in particular, is the atmospheric circulation. An obvious question to pose is how does the influence of circulation alter otherwise simple ideas about cloud and water vapor feedbacks? Studies like Zhang et al. (1994) and Colman et al. (2001) provide hints at the answer to this question and analysis of observations contrasted with model results discussed in section 9 further suggests that these effects cannot be dismissed a priori. To emphasize this point further, the simple cloud liquid water path–optical depth feedback is now reviewed.

Paltridge (1980) first introduced the idea of a cloud optical depth feedback. He proposed that a feedback might exist given the association between optical depth and liquid water path introduced first by Stephens (1978) and given that the relation between liquid water path and temperature was controlled in a manner predicted by simple thermodynamic relationships (as examined later by Betts and Harshvardhan 1987). The early notions of this feedback were posed for a very simple climate system

$$R_{\text{TOA}} = \frac{Q_o}{4} \{1 - \alpha[\text{LWP}(T_s)]\} - \sigma T_p(T_s)^4, \quad (23)$$

which is merely a simplification of (16) with $X_1 = \text{LWP}(T_s)$ and other feedback mechanisms ignored. Paltridge (1980) and later Somerville and Remer (1984) suggested that this system might apply to regions of extensive boundary layer clouds and to the global climate system as a whole given the understood broad importance of these clouds to the global energy budget. It follows from (20) and (23) that the feedback parameter is

$$f = -\frac{1}{\lambda_o} \frac{Q_o}{4} \frac{\partial \alpha}{\partial \text{LWP}} \frac{d\text{LWP}}{dT_s}. \quad (24)$$

Somerville and Remer (1984) use cloud physics data in an attempt to quantify the factor $d\text{LWP}/dT_s$, which they express in terms of the quantity

$$\mathcal{F} = \frac{1}{\text{LWP}} \frac{d\text{LWP}}{dT_c},$$

where T_c is the cloud temperature and where it is implicitly assumed that this sensitivity is identical to the factor in (24), namely,

$$\mathcal{F} = \frac{1}{\text{LWP}} \frac{d\text{LWP}}{dT_c} = \frac{1}{\text{LWP}} \frac{d\text{LWP}}{dT_s}. \quad (25)$$

Using a single-column RCE model, Somerville and Remer (1984) estimate that $\Delta T_{\text{eq},o} \sim 1.74$ for a doubling of CO_2 without feedback (i.e., $\mathcal{F} = 0$) and $\Delta T_{\text{eq},f} \sim 0.75$ for a system with feedback specified by $\mathcal{F} = 0.05$, which is a value that reasonably represents the cloud

data they considered. From (6), (11), and (17), the ratio of these responses follows as

$$\frac{\Delta T_{\text{eq},f}}{\Delta T_{\text{eq},o}} = \frac{1}{(1-f)}$$

and $f = -1.3$ implying a strong negative feedback.

The question surrounding this feedback concept concerns the extent to which such a simple feedback operates in the real climate system? Analyses that correlate global observations of cloud optical depth and cloud LWP data with cloud temperature (Tseloudis and Rossow 1994; Greenwald et al. 1995) show convincingly that the cloud liquid water path–temperature relationship is grossly affected by factors other than temperature (Fig. 11). DelGenio and Wolf (2000) analyzed cloud observations collected at a single site and reached a similar conclusion arguing that the change in LWP observed at that site occurs more through cloud thickness changes than through basic thermodynamic effects. The authors of these studies argue that atmospheric circulation exerts an overriding influence on the distribution of LWP and establishes, to first order, the observed relationship between LWP and temperature. The GCM model analysis of Colman et al. (2001) suggests that even if the feedback were to operate in the simple fashion proposed by Paltridge, its strength is most likely small. This conclusion is a consequence of the large water contents of the clouds in the model analyzed by Colman et al. (2001), producing only marginal changes to the cloud albedos as the water content is increased. Colman et al. (2001) also argue that this feedback is strongest for low water content clouds like cirrus, a point noted in earlier RCE studies of Stephens and Webster (1981) and in later GCM studies (e.g., Roeckner 1988).

d. Cloud feedbacks and precipitation

It was mentioned above that, to first order, the energy balance of the global atmosphere occurs largely as a balance between the radiative cooling of the atmosphere and latent heating associated with precipitation. Cloud feedbacks that affect the radiative heating of the atmosphere will also influence the response of the hydrological cycle to any climate forcing. To examine these ideas, consider the global-mean atmospheric energy budget:

$$R_{\text{atm}}[\epsilon, T, X(T), \dots] = F_{\text{SH}} + F_{\text{LH}}, \quad (26)$$

where R_{atm} represents the net radiation budget of the atmosphere and F_{SH} and F_{LH} are the fluxes of sensible and latent heating, respectively, at the atmospheric–surface interface. Here F_{LH} is also given as

$$F_{\text{LH}} = LP, \quad (27)$$

where L is the latent heat of vaporization and P is the global-mean surface precipitation. For the sake of discussion, consider the system defined by (26) with a

single feedback established by the process $X(T)$. Under the influence of a small climate forcing induced by the action $\Delta\epsilon$, the perturbation energy balance is

$$\Delta R_{\text{atm}} \approx L\Delta P, \quad (28)$$

where perturbations to the sensible heat term are ignored and

$$\Delta R_{\text{atm}} = \frac{\partial R_{\text{atm}}}{\partial \epsilon} \Delta\epsilon + \left(\frac{\partial R_{\text{atm}}}{\partial T} + \frac{\partial R_{\text{atm}}}{\partial X} \frac{dX}{dT} \right) \Delta T. \quad (29)$$

As in Allen and Ingram (2002), we write (29) as follows:

$$\Delta R_{\text{atm}} = \Delta Q_{\text{atm}} + \Delta R_T = L\Delta P, \quad (30)$$

where

$$\Delta Q_{\text{atm}} = \frac{\partial R_{\text{atm}}}{\partial \epsilon} \Delta\epsilon \quad (31)$$

is the radiative forcing of the atmosphere. For example, a doubling of CO_2 decreases the net (upward) TOA flux by $3\text{--}4 \text{ W m}^{-2}$ (depending on whether the TOA is considered to be the tropopause or the true TOA). The net (upward) surface flux is also decreased by about 1 W m^{-2} (e.g., Ramanathan 1981) leaving $\Delta Q_{\text{atm}} \sim 2\text{--}3 \text{ W m}^{-2}$. Thus in the absence of any change in temperature and without related feedbacks, increasing CO_2 slightly decreases the net atmospheric cooling.

The second term of (30), namely,

$$\Delta R_T = \left(\frac{\partial R_{\text{atm}}}{\partial T} + \frac{\partial R_{\text{atm}}}{\partial X} \frac{dX}{dT} \right) \Delta T, \quad (32)$$

is the component of the perturbed budget that depends on T directly and through the feedback process $X(T)$. In principle, the latter process may differ from the set of processes that affect the TOA budget. Specifically, the effect of clouds on the atmospheric column-integrated solar heating is small by comparison to the effect of clouds on the IR column cooling (e.g., Fig. 7 and related discussion) and in marked contrast to the effects of clouds on TOA solar fluxes.

It is convenient to separate the clear-sky contribution $\Delta R_{\text{clr},T}$ and the atmospheric cloud radiative forcing ΔC_T such that

$$\Delta R_T = \Delta R_{\text{clr},T} - \Delta C_T$$

and the clear-sky component of this budget is proportional to column water vapor content, w , which, in turn varies with SST in a quasi-exponential manner resembling the Clausius–Clapeyron (C–C) relation (Stephens 1990; Stephens et al. 1994). From the observed relationship between clear-sky $R_{T,\text{clr},w}$ and SST (Fig. 12), and for small variations of SST, the following simple relation is proposed:

$$\Delta R_{T,\text{clr}} \approx K_T \Delta \text{SST}, \quad (33)$$

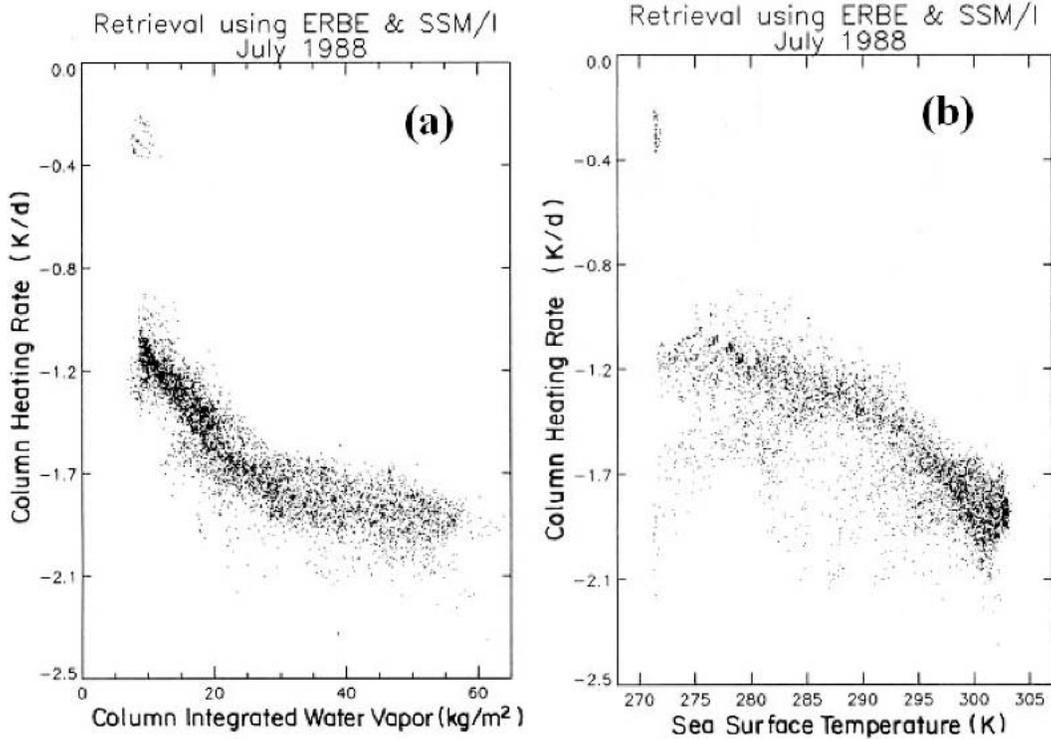


FIG. 12. (a) The relationship between longwave component of R_{atm} and column water vapor. (b) Same as in (a) but for SST (from Stephens et al. 1994).

where $K_T = 1.88 \text{ W m}^{-2} \text{ K}^{-1}$ for $\text{SST} < 290 \text{ K}$ and $K_T = 6.6 \text{ W m}^{-2} \text{ K}^{-1}$ for $\text{SST} > 290 \text{ K}$ broadly brackets the changes in clear-sky water vapor emission. The sharp increase in column cooling above 290 K implied by these values reflects the effects of the supergreenhouse over the warmer, moist tropical atmosphere (e.g., section 5). Combining (33) and (30), we now obtain

$$\Delta Q_{\text{atm}} - \Delta C_T + K_T \Delta T = L \Delta P. \quad (34)$$

We return to this expression below in discussion of the analysis of CMIP data.

7. GCMs and the cloud parameterization problem

A common thread throughout much of the discussion thus far is the importance of the atmospheric circulation to the cloud feedback problem. Since GCM climate models resolve these large-scale atmospheric motions, these models are essential tools in the study of cloud feedbacks. However, most of the cloud processes shaped by atmospheric motions considered relevant to climate feedback occur on scales smaller than typically resolved by these models. Approximations that have to be developed to represent these processes are referred to as the “cloud parameterization” problem and are pursued along two essentially separate paths.

- *Convective parameterization:* The scale of convective clouds lies below the native resolution of the model and the parameterization of convection contains much empiricism. The subgrid-scale parameterizations that deal with these unresolved clouds primarily focus on representing the effects of convection in drying and warming the large-scale atmospheric environment. These empirical convective parameterizations produce the majority of the precipitation “predicted” by most climate models, especially at low latitudes.
- *Large-scale parameterization:* In this case the cloud properties are parameterized in terms of the thermodynamic and dynamical fields resolved by the model. One of the main functions of these parameterizations is to serve as input to the calculation of the model radiation budget. Large-scale clouds also produce precipitation although typically much less than convective precipitation.

A brief history of cloud parameterization

The cloud parameterization problem embodies the all-important return part of the cloud feedback loop (Fig. 10b) in GCMs. Progress on cloud–climate feedback hinges on progress on cloud parameterizations in GCMs. Here is a brief history of the treatment of clouds in GCMs.

- *The 1960s:* In this period, clouds were prescribed in zonal average form and then primarily in terms of their areal amount and height (high, middle, and low cloud). Clouds only interacted with radiation, albeit

crudely, and played no explicit role in the hydrological cycle of the model.

- *The 1970s:* During the early years, the treatment of clouds remained largely unchanged from that of the previous decade although the introduction of nonzonally averaged cloudiness began to emerge. More sophisticated treatment of cloud–radiation processes became available in the latter part of this decade with studies that indicated how key cloud optical properties could be related to the amount of condensed water and ice in clouds. A cloud prediction scheme based on cloud physical principles, expressed in terms of these cloud water and ice contents (hereafter cloud condensate), emerged with the study of Sundquist (1978) almost 20 years before these schemes (referred to as prognostic schemes) would begin to systematically replace the empirical diagnostic methods in vogue during this period. Another study ahead of its time was that of Twomey (1977) who demonstrated an effect of aerosol on cloud albedo (later to be known as the Twomey effect). This effect currently occupies a prominent role in cloud–climate research. As in the previous decade, cloud feedback concepts began to emerge in a simple form expressed in terms of cloud amount and cloud-height properties (e.g., Schneider 1972) and the relation of these properties to surface temperature.
- *The 1980s:* Diagnostic cloud schemes continued to be the common way of deriving clouds in models and some of these schemes introduced cloud condensate via empirical relationships in an attempt to capture a more realistic interaction between radiation and clouds. A few global models adopted prognostic cloud schemes (e.g., Smith 1990). Cloud feedback concepts of this period, such as the cloud water and ice content–temperature feedbacks explored in the simple RCE studies discussed previously were also explored in GCMs (e.g., Roeckner 1987; Mitchell et al. 1987; Le Treut and Li 1991; and others) yielding results similar to the earlier RCE studies. The need to include interactive cloud–radiation processes as opposed to fixed processes in GCMs began to be more fully appreciated by the modeling community. Despite these advances though, cloud schemes only produce “large scale” cloudiness that interacted with the radiation budget and the latter remained essentially empirically connected to the bulk of the model’s hydrological cycle. The complex nature of cloud feedback and the intimate coupling of these feedbacks to other climate feedbacks, such as surface albedo feedbacks, also emerged.
- *The 1990s:* Diagnostic cloud schemes were gradually replaced by prognostic condensate schemes albeit with crude “bulk” representations of microphysics and the importance for including interactive cloud–radiation processes continued to be underscored. The cloud physics, however, remained largely decoupled from the treatment of convection and model precipi-

tation. Emphasis began to shift toward regional and cloud-resolving models that represent cloud and precipitation processes in a more physically consistent and explicit manner (e.g., Browning 1993). Cloud feedbacks in global models were shown to be highly uncertain (e.g., Cess et al. 1990), sensitive to the details of cloud prediction and the way radiation was coupled to cloud properties (e.g., Senior and Mitchell 1993). Cloud feedbacks dealing with the coupled ocean–atmosphere system also emerged (Ma et al. 1994) only to suggest an even more acute sensitivity of the coupled system to cloud feedbacks. Focus began to shift toward more detailed cloud microphysical processes and there was a growing realization that aerosol can affect cloud feedback through the Twomey effect.

- *The present:* Interest in the effects of aerosol not only on cloud radiative processes but also on precipitation-forming processes (e.g., Rosenfield 1999) brings to the forefront a continuing emphasis on microphysics. The use of explicit cloud process models imbedded in global models has now emerged (Grabowski 2001; Randall et al. 2003) with the intent on a more self-consistent treatment of clouds and radiative processes as an integral part of the hydrological cycle.

8. Cloud feedbacks in GCMs

a. Sensitivity to cloud parameterization

It is clear from many GCM studies that the sensitivity of GCM climate models depends on the way clouds are parameterized, including details of the ice cloud physics and their radiative properties (Fowler and Randall 1994; Ma et al. 1994), cloud overlap methods (Liang and Wang 1997) as well as the atmospheric cloud flux (e.g., Slingo and Slingo 1988; Randall et al. 1989; and others). What emerges from many of these studies is a repeated demonstration of the importance of dynamics through its influence on the vertical organization of cloud properties affecting the profile of radiative heating and the influence of this heating profile on the model's hydrological cycle through the indirect influence on convection.

In a more systematic evaluation of the sensitivity of a model to cloud parameterization, Mitchell et al. (1987) and later Senior and Mitchell (1993) examined the response of a GCM model to a doubling of CO₂ when different versions of a cloud parameterization were employed in the same model. The results of their studies are summarized in Fig. 13 showing a sequence of global warming predictions as different feedbacks are systematically introduced. The first estimate corresponds to the warming in the absence of feedback, the next with water vapor feedback added, the third with snow/ice albedo feedbacks added to water vapor feedback and finally the case when cloud feedback is added. Senior and Mitchell (1993) found that the presence and ab-

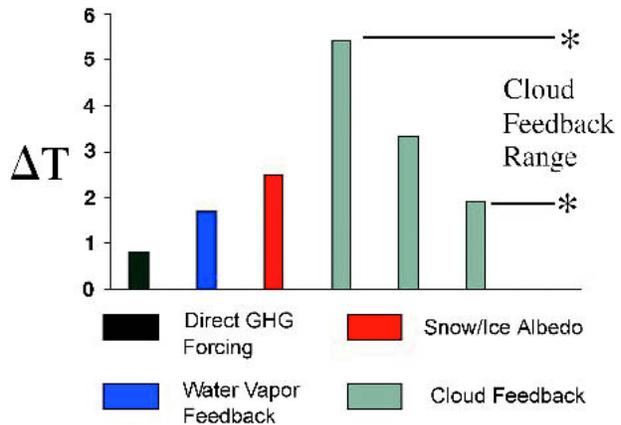


FIG. 13. The response of a single climate model to an imposed doubling of CO₂ as different feedbacks are systematically added in the model (adapted from Senior and Mitchell 1993). Different treatments of cloud processes in the model produce a large spread in predicted surface temperature due to CO₂ doubling.

sence of microphysical and optical thickness feedbacks, which were permitted in different versions of the parameterizations, produce a range in warming between 1.9° and 5.4°C. Yao and Del Genio (2002) also reported that the 2 × CO₂ sensitivity of the Goddard Institute for Space Studies (GISS) GCM was similarly dependent on the details of the moist convection and large-scale cloud parameterizations.

b. GCM intercomparison studies

The specific differences between sensitivities of different climate models shown in Fig. 1 has prompted a number of model intercomparison studies over the years. Although these studies point to the differences in cloud–radiation parameterizations as a principal factor in the spread of the model responses, these studies unfortunately have not provided enough information about the in-cloud properties to shed much light on how specific details of different parameterization formulations affect cloud feedbacks in the models.

1) THE ±2 K SST PERTURBATION INTERCOMPARISONS

Quantifying feedback in complex, coupled model systems is complicated and, in principle, requires comparison between separate model integrations with and without the feedbacks and with and without the climate forcing (e.g., Schlessinger 1988; Wetherald and Manabe 1988). From the perspective of the community intercomparison efforts, such an approach to feedback analysis is problematic. Parallel simulations (a minimum of four) require computational resources that deters broad participation in community intercomparison efforts.

Cess and Potter (1988) introduced a simple and valu-

able experiment concept in an attempt to assess the accumulated effect of feedbacks in GCMs. The procedure required just two model integrations with climatological July SST distributions perturbed by ± 2 K. By further restricting analysis to 60°N – 60°S , influences of snow/ice feedbacks are minimized. Fixing the SSTs also removes the complicating effects of the dependence of the model response on its control climate. The procedure outlined by Cess and Potter (1988) proposes short integrations of 105 days with analysis performed on averages over the last 30 days.

Given the fixed SST boundary conditions of the ± 2 K SST experiments, all models produce approximately the same global-mean surface temperature response although the actual responses differ slightly due to model-to-model differences in the land temperature responses. Cess et al. (1990) propose that the difference in radiation imbalance at the top of atmosphere at the end of these short integrations (i.e., ΔR_{TOA}) is a measure of the forcing. However this imbalance not only contains combinations of the initial forcing induced by the instantaneous changes to SST but also the effects of feedbacks that occur on relatively short time scales. To separate the contribution of feedbacks from this response requires both an estimate of the forcing as well as an estimate of the sensitivity of the equivalent open system.

Expressing this net response in terms of its clear- and cloudy-sky components

$$\Delta R_{\text{TOA}} = \Delta R_{\text{TOA,clr}} - \Delta C_{\text{TOA,net}},$$

then simple rearrangement produces

$$\frac{\Delta R_{\text{TOA,clr}}}{\Delta R_{\text{TOA}}} = 1 + \frac{\Delta C_{\text{TOA,net}}}{\Delta R_{\text{TOA}}}, \quad (35)$$

which is interpreted by Cess et al. (1990) as a direct measure of cloud feedback. This interpretation follows if we consider the quantity $-\Delta R_{\text{TOA}}/\Delta T$ as equivalent to the sensitivity λ_f defined in (11) and further that the quantity $-\Delta R_{\text{TOA,clr}}/\Delta T$ is equivalent to λ_o . With these assumptions

$$\frac{\Delta R_{\text{TOA,clr}}}{\Delta R_{\text{TOA}}} = \frac{\lambda_o}{\lambda_f} = \frac{1}{1-f}, \quad (36)$$

from which we obtain

$$\frac{\Delta C_{\text{TOA,net}}}{\Delta R_{\text{TOA}}} = \frac{f}{1-f},$$

implying that the magnitude of the cloud feedback, f , is directly related to the cloud net flux $\Delta C_{\text{TOA,net}}$.

Figure 14a shows the distribution of the quantity $(\Delta R_{\text{TOA}}/\Delta T)^{-1}$ as a function of $\Delta C_{\text{net}}/\Delta R_{\text{TOA}}$. Given the ranges of this quantity in Fig. 14a, we infer that f varies between 0.64 and -5.3 . Cess et al. (1996) provided an update on their 1990 analysis with newer versions of the models used in the earlier study. In this update, the range in response narrowed (Fig. 14b) be-

tween models, but it was shown that the responses were arrived at in very different ways from model to model (Fig. 14c) with considerable model-to-model variations in the long- and shortwave contributions.

This simple analysis suggests that cloud feedbacks in models are highly variable from model to model. Although the experiments proposed proved a useful and convenient framework for the study response of different models to a fixed forcing, the quantitative interpretation of these results in terms of feedbacks specifically has turned out to be misleading for a number of reasons:

- 1) The system represented in these experiments is not in radiative equilibrium, a necessary assumption in arriving at (6) and all subsequent feedback analysis.
- 2) Derived at the end of the integration ΔR_{TOA} is the net response of the system and not its forcing and ΔC_{net} is similarly the cloudy-sky portion of this net response. If as usually considered, the forcing is taken to be the perturbation to the radiative budget derived immediately after the SSTs are changed, then the sign of the forcing is the reverse of that assumed in the original analysis. This suggests that the sign of the feedbacks too are different, a point noted by Soden et al. (2004).
- 3) Also, $\Delta R_{\text{TOA,clr}}$ is not a measure of the system response without cloud feedbacks and thus does not relate to λ_o in any obvious way. Because clouds affect the clear-sky portions of the radiation budget, it is not possible to use this clear-sky component of the budget to identify the system without cloud feedback and this is a source of misrepresentation of cloud feedback in the analysis of these ± 2 K SST experiments (Colman 2003; Soden et al. 2004).
- 4) When different assumptions are made to define the “system” then very different estimates of feedback are obtained. Arking (1991) uses the data of Cess et al. (1990) and employs different assumptions about the nature of the system to arrive at values of f ranging from -0.25 to 0.45 , very different from the range of Cess et al. (1990). Soden et al. (2004) note that the feedbacks diagnosed using the method of Cess et al. (1990) differ not only in magnitude but generally in sign from estimates of the feedback derived using the partial radiative perturbation method.

2) CMIP

The sensitivity of coupled ocean–atmospheric models to cloud feedbacks differs from the sensitivity of models with fixed SSTs (e.g., Ma et al. 1994; Williams et al. 2003; and also acknowledged in Cess and Potter 1988). CMIP (Meehl et al. 2000), under the auspices of the World Climate Research Program (WCRP) Working Group on Coupled Models (WGCM) was created specifically to examine and compare coupled models forced by $1\% \text{ yr}^{-1}$ increases in CO_2 .

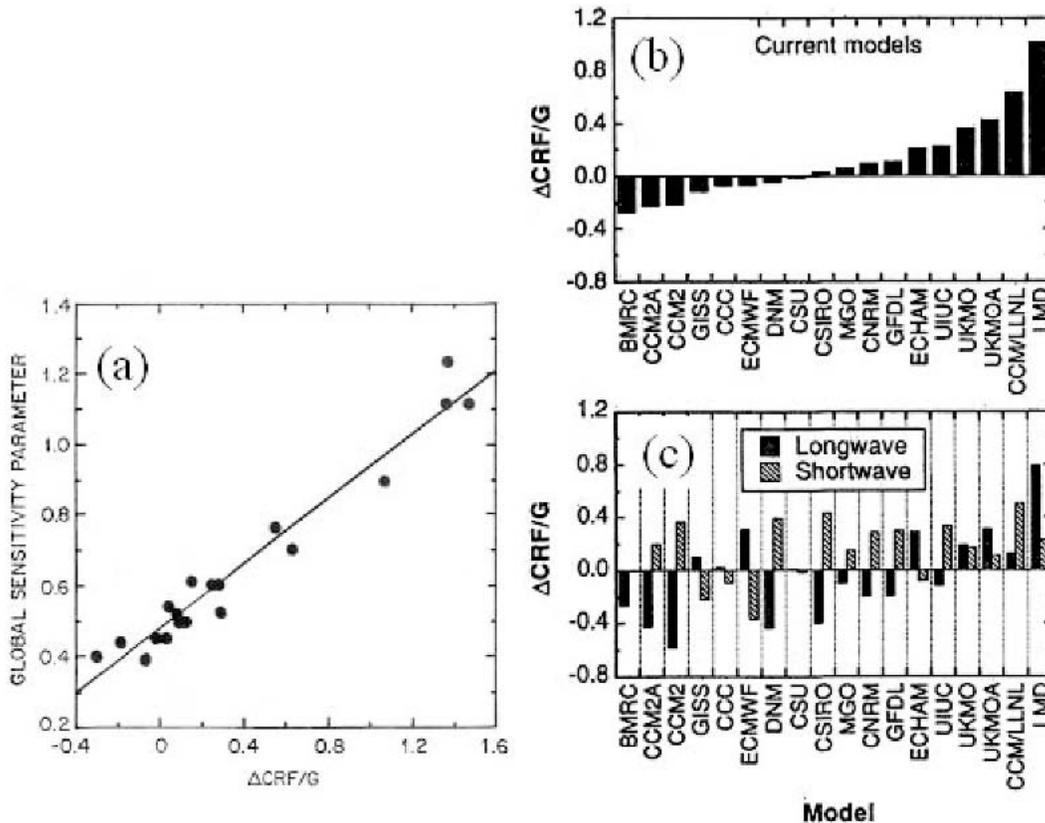


FIG. 14. (a) The global sensitivity parameter as defined by Cess for the ± 2 K SST perturbation experiments plotted as a function of the ratio of the change in C_{net} to change in net radiation budget for 19 GCMs. (b) This ratio derived from updated group of models in the Cess et al. (1996) update. (c) The breakdown of long- and shortwave components of the ratio.

Results from CMIP are presented in Figs. 15a,b in a manner that illustrates the relationship between atmospheric radiation balance and precipitation described by both (28) and (34). All results are expressed in terms of globally averaged quantities and averaged over the last 20 yr of the 80-yr CMIP integrations. This broadly corresponds to the time of an approximate doubling of CO_2 from its initial state. The quantities plotted are the differences between the model integrations with and without the prescribed forcing. Figure 15a presents the change in net flux at the surface correlated with the change in precipitation $L\Delta P$. The surface net flux difference is, for all intents, equivalent to ΔR_{atm} since the net change in TOA is practically zero. These results support the earlier conjecture that the precipitation and column cooling are highly correlated on the global annual space-time scale. The results also imply that changes to the intensity of the hydrological cycle are controlled by the perturbations to the energetics of the climate system.

Figure 15b contrasts the relationships between the difference in both the net flux (ΔR_{atm}) and $L\Delta P$ as a function of the predicted change in surface air temperature ΔT . The shaded portion of the diagram represents

the range of $\Delta R_{atm,clr}$ that is expected given (30) assuming the two values of K_T inferred from Fig. 12. These results indicate that changes to the net atmospheric radiative budget, in response to a climate forcing, do not simply follow projected changes in the clear-sky radiative cooling. The changes to the net radiation budget, and the equivalent changes to precipitation, $L\Delta P$, also do not obviously correlate well with the predicted warming exhibiting a spread between models. Since clouds are a principal modulator of the atmospheric radiative heating, it is likely that the scatter in both the predictions of warming (Fig. 1) and changes in global precipitation (Figs. 15a,b) are a consequence of different cloud feedbacks in the models.

9. Evaluating models

To date, cloud feedback studies have followed typically one of two paths. One uses observations primarily in an attempt to diagnose the key mechanisms of the feedbacks. These studies, however, are invariably inconclusive for reasons already described. Unless rare circumstances exist in which nature provides a way of

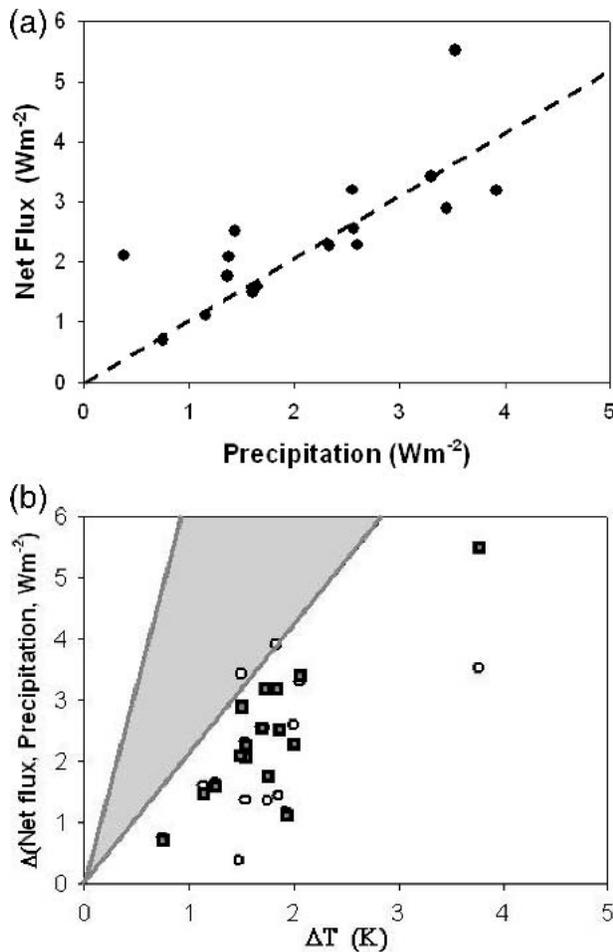


FIG. 15. (a) The relationships between the change in global-mean precipitation and the surface net flux. (b) The change in the globally averaged precipitation (open circles) and net surface fluxes (squares) as a function of the change in surface air temperature. The shaded area adjusted downward by (ΔQ_{atm}) represents the clear-sky cooling change predicted by (36) with the two given values of K_7 . All quantities in (a) and (b) are the differences between the 20-yr average (1961–80) of a 1% yr⁻¹ increase of CO₂ (thus roughly corresponding to a doubling of CO₂) and the parallel simulation with fixed CO₂. Data are from CMIP (courtesy of B. Soden 2003, personal communication).

observing the climate system with key processes turned off (Soden et al. 2002), then feedbacks per se cannot generally be diagnosed directly from observations alone without resorting to unrealistic assumptions about the “climate system” and its open and closed forms. The second path of study primarily relies on models, the complexity of which vary considerably from study to study. While it is possible to identify feedbacks in model systems, the tools we use to do so are coarse and not mature. Without detailed evaluation of these models, we cannot be confident about their relation to reality. Progress toward understanding cloud feedback requires a more definitive use of observations and the development of diagnostic methods

that provide a more stringent evaluation of models. It is also reasonable to suppose that a necessary test of a climate model is the requirement to reproduce the observed present-day distributions of clouds, their effects on the earth’s energy budgets, and their relation to other processes, as well as be able to reproduce observed climate variability. Recently, a number of diagnostic studies have proposed ways to identify objectively cloud regimes that perhaps offers a useful framework for evaluating cloud processes in models.

a. Assessment of cloud occurrence

Perhaps the most basic test of model parameterization of clouds is to quantitatively evaluate whether models place clouds in the atmosphere at the same time and place as observed. Explicit evaluation of cloud occurrence is possible using the cloudiness predicted by a weather forecast model at given instances in time and at given locations in space compared to observed clouds at the same place and time. This kind of evaluation might also be possible using a climate initialized with updated analysis fields and run in weather prediction mode.

Two examples of this type of study are provided by Miller et al. (1999) and Hogan et al. (2001). The Miller et al. (1999) study matched the 24-h forecasts provided by the European Centre for Medium-Range Weather Forecasts (ECMWF) to 11 days of lidar profiles obtained by the lidar system flown on the space shuttle in 1996 as part of the Lidar In space Technology Experiment (LITE; Winker et al. 1996). The results of this study indicated that the vertical locations of clouds and alternatively clear-sky layers matched those identified by LITE for about 75%–90% of the time. Similar kinds of results were obtained by Hogan et al. in their comparison of ECMWF cloud prediction with surface radar data compiled from one site over many months. Although these studies indicate that clear biases exist between the model and observations, hinting at biases in critical cloud processes in the model, these studies highlight the realism in the occurrences of cloudiness predicted by the forecast model underscoring the potential of these models in the study of cloud feedback.

b. Properties of cloud regimes

Comparing the cloud occurrence statistics of models with observations is a necessary test of cloud prediction but it is important to place these kinds of comparison within the context of other properties of clouds and their environment. This task can be reduced to a manageable scope by realizing clouds organize themselves into a smaller subset of regime types. This is the approach of the Global Energy and Water Cycle Experiment (GEWEX) Cloud System Study (GCSS; Browning 1993) and is illustrated in the recent study of Jakob and Tselioudis (2003) who perform a simple cluster analysis of tropical western Pacific ISCCP histogram data. In defining the ISCCP histogram as a vector with

42 τ -CTP elements, their cluster analysis identified four main cloud regimes that represent the majority of the cloudiness of that region. The results of their analysis are summarized in Table 2. Their analysis also reveals the prevalence of shallow clouds and in applying the cluster method to cloudiness derived from the ECMWF forecast model, the comparison with ISCCP (also given in Table 2) reveals that the forecast model significantly overestimates both the amount and frequency of these low clouds and their optical depths relative to ISCCP (Fig. 16).

An objective identification of cloud regimes, in principle, provides a strategy for examining other properties of these cloud regimes and, perhaps, the key processes that establish them. Figure 17 is an example of radiative flux data grouped via the cloud regimes of Jakob and Tselioudis (2003). The surface and TOA radiative fluxes shown are derived from measurements made at a Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) site in the tropical western Pacific (Jakob et al. 2005). The flux data are grouped by regime, presented in the form of a box whisker diagram, and the different radiative characteristics of each regime are highlighted.

c. Processes I: Cloud–radiation interactions

Many GCM studies rely on simple, cursory comparison between model and observed climatological global cloud cover and/or TOA cloud radiative forcing as a way of demonstrating the realism of simulations. However, merely reproducing distributions of observed parameters independent of one another is not an adequate test of models since it is possible to tune to the observations using any one of many combinations of cloud parameters that, individually, might be unrealistic. The Jakob and Tselioudis (2003) study described above, as well as others below, underscore how apparent errors in optical depth and the amount of any one cloud type can either offset each other when calculating TOA fluxes or be masked by errors in properties of other cloud types.

Simple comparisons of model and observed cloud parameters does not provide any insight into the realism

of those processes essential to feedback. On the other hand, evaluation of *processes*, as opposed to cloud *parameters*, requires assessment of key relationships, examples of which are presented in section 6. For example, examination of radiative transfer processes requires an investigation of the relationship between cloud radiative forcing and the various cloud optical and physical properties identified above in section 3 to provide some idea about the cloud radiation processes that are important to cloud feedback. The study of OBH is one, albeit limited, example that attempts to relate the amounts of different cloud types to the radiative forcing. Webb et al. (2001) use this same analysis approach in an attempt to examine the nature of the differences in cloud radiative processes implicit in the ERBE cloud forcing data and the forcings derived from three global models.

Figure 18 presents the key results of Webb et al. (2001) for the TWP region where the data are expressed as monthly means for July 1988. Figure 18a is the ISCCP histogram of frequency of occurrence indicating that the most commonly retrieved cloud tops in this region are in the upper troposphere with optical thicknesses ranging from thin to thick clouds, a result already evident in the cluster analysis of Jakob and Tselioudis (2003). Figure 18b presents the contributions to both C_{SW} and C_{LW} (in $W m^{-2}$) by high (tops above 440 hPa), middle (tops between 680 and 440 hPa) and low (tops below 640 hPa) clouds as defined by ISCCP (in yellow) as well as the percentage of occurrence of each cloud type for that month (in green). The total short- and longwave cloud fluxes derived from ERBE are shown below the abscissa in red for comparison. The fact that high clouds contribute most to the cloud fluxes of this region basically reproduces the earlier findings of OBH. The remaining panels present the same data from the three global models.

For sake of discussion, consider the results from the ECMWF model (Figs. 18e,f). This model produces slightly lower amounts of high thin and thick clouds but more optically thick boundary layer cloud than retrieved by ISCCP (Figs. 18a,b) also consistent with the findings of Jakob and Tselioudis (2003). The value of C_{SW} summed from each cloud type is, however, close to the ERBE-observed value although the total C_{LW} is not. This analysis shows how decomposing the cloud fluxes into its high, middle, and low and thin to thick cloud components reveals how model biases in one cloud type can be offset by biases of other types. An understanding of why these differences occur requires a deeper level of comparison of model and observed cloud physical and optical properties than can be typically gleaned from current global observations.

d. Processes II: Clouds and dynamics

According to the control theory view of feedback systems, feedbacks are defined by relationships of the

TABLE 2. The frequency of occurrence (frequency) statistics and total cloud amount (TCC) (both in %) for five categories of clouds derived from ISCCP data and ECMWF forecast data for the TWP region mentioned in the text. The histograms corresponding to the first three of these categories are presented in Fig. 17.

Cluster category	ISCC frequency	ISCCP TCC	ECMWF frequency	ECMWF TCC
Cluster 1	33	32	59	53
Cluster 2	33	75	21	92
Cluster 3	11	99	14	99
Cluster 4	17	94	6	99
Cluster 5	6	76	1	90

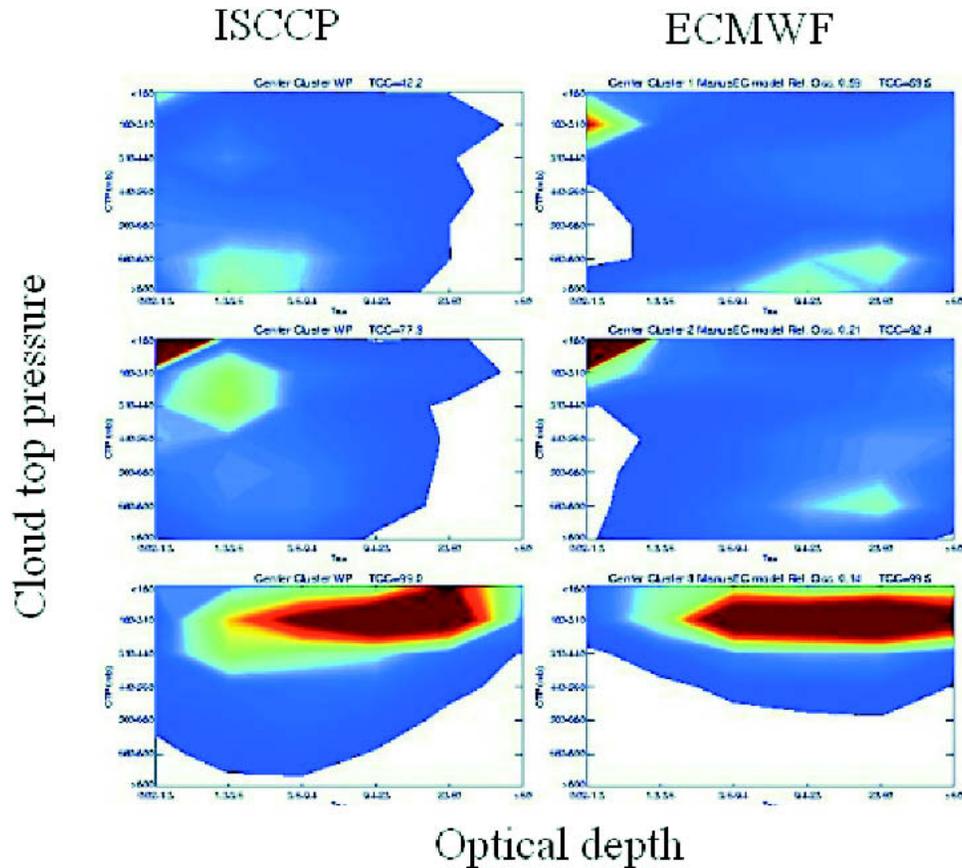


FIG. 16. CTP- τ histograms of the centroids of a three out of a five cluster analysis of cloud data for 1999 for the TWP (10°N – 10°S , 130° – 170°E ; Jakob and Tselioudis 2003). The five regimes identified by this cluster analysis are 1) mainly shallow clouds, 2) some deep convective clouds but a dominance of midlevel clouds, 3) mostly transparent cirrus, 4) a convective regime with large amounts of high-level clouds, and 5) large amounts of deep high-top clouds. The histograms for the first three of these are shown in order from top to bottom.

type $X(T)$ that relate a cloud parameter X to the system output variable T . It has been argued that imbedded in this relationship is an important pathway involving large-scale circulation. Recent studies attempt to examine the nature of these relationships comparing correlations of model quantities with the same correlations derived from observations. Most studies use reanalysis data derived from the numerical weather prediction (NWP) model and analysis systems, and much of the focus has been directed to extratropical weather systems for one good reason—the dynamical forcing of these systems tends to be large and well predicted by the NWP model. Studies such as Norris and Weaver (2001), Weaver and Ramanathan (1996, 1997), and Del Genio and Kovari (2002) among others follow the approach of Bony et al. (1997) and use of 500-hPa vertical motion fields derived from reanalysis as the key index of the dynamics. The reanalysis vertical motion is however problematic since the quality of these particular data is generally considered suspect especially in tropi-

cal regions where the large-scale dynamical forcings are generally weak.

1) MIDLATITUDE BAROCLINIC SYSTEMS

An important example of the composite method is the study of Lau and Crane (1995). This study relates the cloudiness of synoptic-scale midlatitude baroclinic weather systems to the general dynamical features of these systems. Composites were formed from 10 years of meteorological data obtained from the ECMWF reanalysis and cloud information from ISCCP. This study was later extended by Klein and Jacob (1999) who compared the ECMWF-predicted cloud distributions to those of ISCCP. The distributions of high, middle, and low clouds obtained from a composite of all such systems over the North Atlantic were superimposed on the composite 1000-hPa height field and surface wind anomalies obtained from the ECMWF analyses (Fig. 19a). Figure 19b is the equivalent analysis applied to

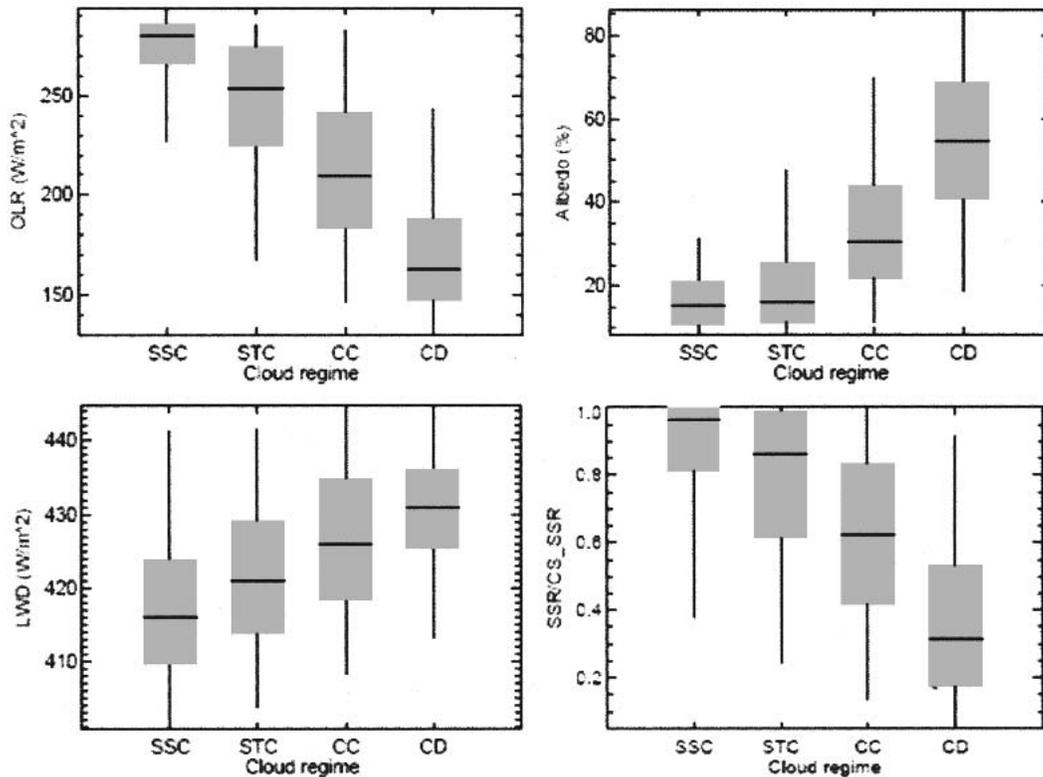


FIG. 17. Box whisker representation of the radiative fluxes measured at the Manus ARM site as a function of cloud regime. (top left) OLR and (top right) visible TOA albedo and (bottom left) downward surface long- and (bottom right) shortwave fluxes, the latter being the ratio of cloudy to clear-sky fluxes (bottom two panels). The horizontal line is the median value, the extents of the shaded boxes are the 25th and 75th percentiles, and the max/min values are indicated by the extents of the vertical lines (Jakob et al. 2005).

ECMWF clouds. Figure 19c is a repeat of Fig. 19b but accounting for the transparency of the thin upper-level clouds and, in this way, is more comparable to the ISSCP cloud distributions of Fig. 19a.

A number of important features emerge from the comparison of Figs. 19a,b. (i) As in the study of Jakob and Tseloudis (2003), the distributions of clouds associated with circulation patterns implied in the 500-hPa height field are highly coherent suggesting that clouds associated with this type of weather system are organized into repeatable structures or regimes. (ii) The distribution of clouds as forecast by the model is overall similar to the ISSCP observed clouds, which again suggests a broad degree of realism of the clouds in this model. This further implies that the forecast model may be a credible tool for understanding the processes that determine the relation between atmospheric dynamics and clouds, at least for these weather systems. (iii) Differences between predicted and observed clouds do exist, especially in those regions where a mix of high- and midlevel clouds is suspected to occur. In these regions the model tends to produce generally thinner high clouds than observed, exposing more midlevel clouds (Fig. 18c) than observed by ISSCP. The ECMWF

model, like the majority of climate models, predicts clouds as fields of liquid and ice water contents and unfortunately for the case of high clouds, the ice content cannot be adequately constrained by the cloud optical property information available from present-day satellite observations (Stephens 2000).

Other examples of composite analysis of these extratropical weather systems is described in Tselioudis et al. (2000) and Tselioudis and Jakob (2002). Tselioudis et al. (2000) correlate ISCCP cloud types to surface pressure and note that the difference in cloud-type distributions, varying from optically thick precipitating clouds in low pressure regions to shallower clouds in high pressure regions, produce differences in TOA absorbed radiation of 50 W m^{-2} . Tselioudis and Jakob (2002) used the sign of the 500-hPa large-scale vertical velocity as an index of the dynamics of these systems and contrasted the relationships derived using climate model data and the ECMWF forecast model data together with ECMWF Re-Analysis (ERA) matched to ISSCP clouds. They show that both models tended to overestimate the optical depths of clouds and underestimate the cloud amount in the large-scale descent regimes. Like the earlier examples, systematic differences

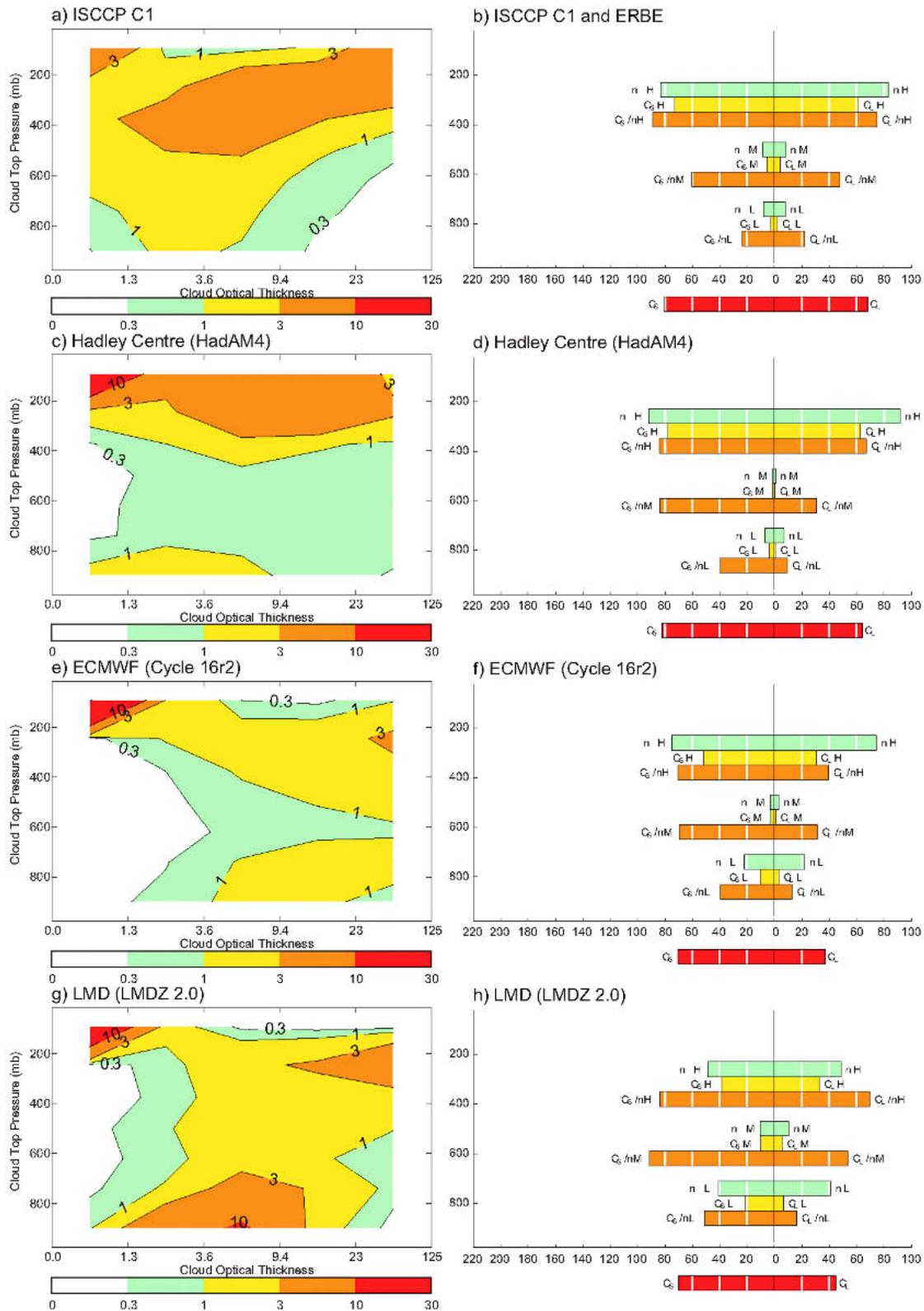


FIG. 18. (a) The Jul 1988 monthly averaged ISCCP-like frequency of occurrence distribution (in %) of the tropical warm pool region. (b) The breakdown of the SW (C_s) and LW (C_l) cloud radiative forcings according to cloud-top pressure. Cloud amounts are indicated by the lengths of the nH, nM, and nL bars. The ERBE cloud forcings are given by the bottom red bar. (c)–(h) The same as (a) and (b) but for the three models (Hadley Centre, ECMWF, and the LMD model). (From Webb et al. 2001.)

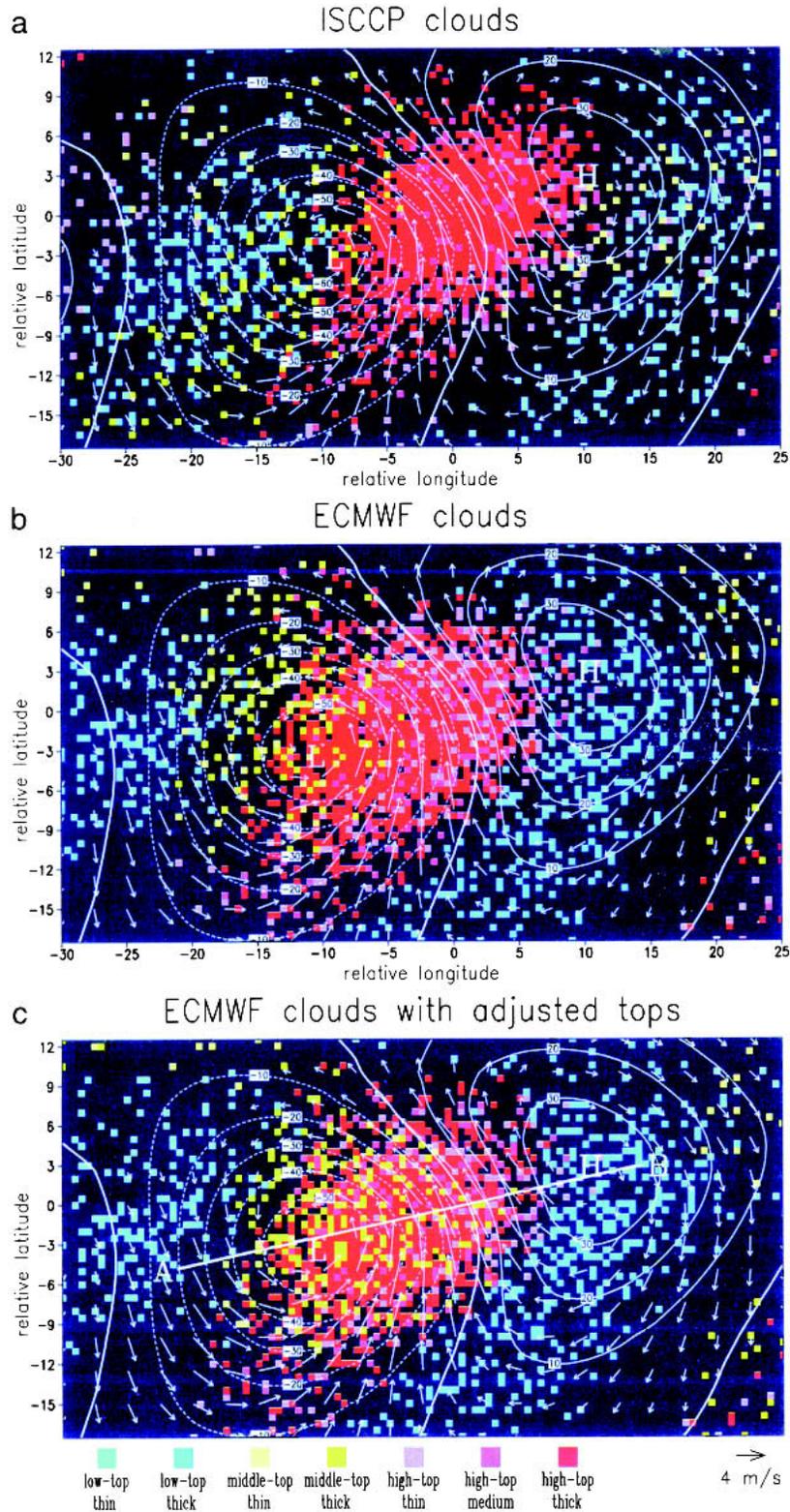


FIG. 19. (a) Distributions of 1000-hPa horizontal wind (arrows) and geopotential height (contours, interval 10 m) from ERA analyses and various cloud types (color pixels) from ISCCP observations as originally presented in Lau and Crane (1995). (b) As in (a) but for the cloud fields from 24-h ERA forecasts. The physical cloud top is used to classify the clouds. (c) As in (b) but with cloud top adjusted for partial cloud transparency (Klein and Jakob 1999).

in these properties cancel each other out when calculating TOA fluxes.

2) TROPICAL CLOUD SYSTEMS

Williams et al. (2003) also employ composite analysis methods to examine the relationships between selected properties of tropical clouds, SSTs, and vertical motion. They examine these relationships within the context of both a forced climate change experiment and natural variability experiments as represented by Atmospheric Model Intercomparison Project (AMIP) style model integrations. The results of these model experiments were obtained with the same version of the Hadley Centre climate model used in the Webb et al. (2001) study.

Williams et al. (2003) define the model response in terms of the quantities $\Delta C_{\text{net}}/\Delta\text{SST}$ and $\Delta N/\Delta\text{SST}$, where N is the cloud amount and Δ refers to a difference between the $2 \times \text{CO}_2$ and the control, averaged over the Tropics and over 5 yr of monthly mean values. They decompose the response of the cloud amount to changes in SST as follows:

$$\frac{\Delta N}{\Delta\text{SST}} = \left(\frac{\Delta N}{\Delta\text{SST}} \right)_{\text{absolute}} + \left(\frac{\Delta N}{\Delta\text{SST}} \right)_{\text{local}} + \left(\frac{\Delta N}{\Delta\text{SST}} \right)_{\text{remote}}, \quad (37)$$

where the first term is the contribution to the total response through changes in cloud amount induced by changes to the tropical mean SST. The second term is the contribution due to effects of local warming/cooling relative to this tropical mean. The third is the contribution by SST changes that occur remotely from the cloudy area under consideration. The remote influences can intuitively be interpreted as relating to gross changes in cloud amount via dynamical processes and thus are perhaps indicative of feedbacks involving dynamics.

Figure 20 presents the Williams et al. (2003) analysis showing the total cloud response according to changes in the SST anomaly ($d\text{SST}$) and changes in vertical velocity $d\omega$. The SST anomaly is derived in such a way as to reflect local changes in SST relative to the tropical-mean SST change and the remote changes through dynamics are supposed to be represented by $d\omega$. The analysis is presented in terms of $d\text{SST}-d\omega$ composites, much in the manner of Bony et al. (1997), and are presented separately for different cloud-type categories (high, middle, and low) and optical depth categories, (thin, medium, and thick).

The results of Fig. 20 indicate a range of variation between different cloud types and thickness and $d\text{SST}$ and $d\omega$. In general, the response of high cloud, and high thick cloud specifically, follows $d\omega$ more so than $d\text{SST}$ as noted by others whereas the reverse is true of low cloud. High thin and medium clouds do show a corre-

lation with $d\text{SST}$ and there is a reduction in high cloud when $d\text{SST}$ is negative and when $d\omega$ indicates increased descent (Figs. 20a–c). Williams et al. (2003) also find that the changes to clouds inferred to be associated with local SST changes are an order of magnitude larger than those induced by absolute changes in SST. This is reflected in the values of the first and second terms of (40), which are provided in the caption of Fig. 20. The cloud response to circulation changes (not given) are also comparable to the responses to local SST changes for some cloud types (K. D. Williams 2003, personal communication).

10. Summary, concluding comments, and outlook

Feedbacks in the climate system associated with clouds continue to be considered as a major source of uncertainty in model projections of global warming. This paper offers a critical discussion of the topic of cloud–climate feedbacks and exposes some of the underlying reasons for the inherent lack of understanding of these feedbacks.

Despite the complexity of the cloud feedback problem, it is argued that the basis for understanding such feedbacks, in part, lies in developing a clearer understanding of the association between atmospheric circulation regimes and the cloudiness that characterizes these “weather” regimes. One of the factors that has limited progress on the topic specifically concerns the problem of the parameterization of cloud processes in global models and the limited evaluation of these representations. While GCM climate and NWP models represent the most complete description of all the interactions between the processes that establish the main cloud feedbacks, the weak link in the use of these models lies in the cloud parameterization imbedded in them. Aspects of these parameterizations remain worrisome containing levels of empiricism and assumptions that are hard to evaluate with current global observations. For example, the relationship between convection, cirrus anvil clouds, and SST is a recurring theme in many feedback hypotheses (section 5) yet the connections between convection and cirrus in parameterization schemes is highly uncertain, in many cases empirical, and difficult to evaluate with observations. This is one area where observations are needed to evaluate cloud parameterization processes and feedbacks derived from these processes.

A second factor that has limited progress concerns the methods developed and used to define and then quantify feedbacks in models. These methods are invariably based on a system’s perspective, the implementation of which has been most problematic.

- 1) The definition of the “control system” is the principal source of confusion in feedback analysis. Perhaps the main limitation of the analysis of feedback, either using observations or GCM output, traces

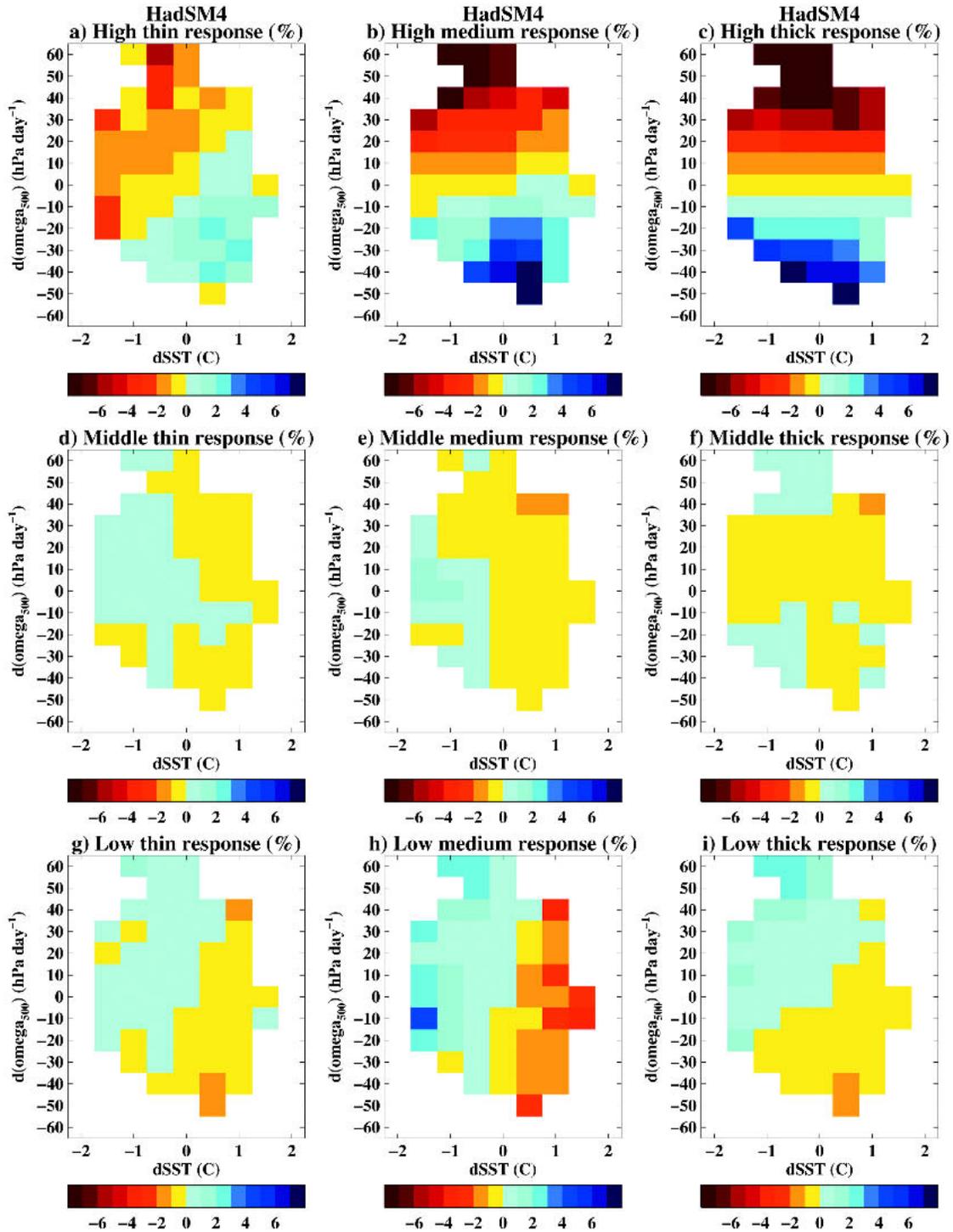


FIG. 20. Change in cloud amount (%) in the Hadley Centre climate model (HadSM4) in response to doubling CO_2 for three ISCCP cloud height and optical thickness ranges. The composites are formed from individual monthly mean grid points binned according to $d\omega_{500}$ and temperature response relative to the mean tropical SST response (dSST). The values given represent the first two terms of the rhs of (40) (modified from Williams et al. 2003) and are as follows: *high thin cloud*, absolute = -0.011 , local = 1.363 ; *high medium cloud*, absolute = -0.004 , local = 0.863 ; *middle thin cloud*, absolute = 0.013 , local = -0.123 ; *middle medium cloud*, absolute = -0.011 , local = -0.511 ; *low thin cloud*, absolute = 0.015 , local = -0.486 ; *low medium cloud*, absolute = -0.013 , local = -1.975 .

back to unrealistic assumptions about the nature of the “system” and its open and closed forms. Out of necessity, most studies make ad hoc assumptions about the overriding importance of one process over all others generally ignoring other key processes, and notably the influence of atmospheric dynamics on cloudiness. Generally little discussion is offered as to what the system is let alone justification for the assumptions given. Much more detail on system and its assumptions are needed in order to judge the value of any study. Problems stemming from the lack of understanding of the system are referred to as the system identification problem (e.g., Bellman and Roth 1983) and diagnostic procedures exist to identify engineering systems when only partial information about them is available.

- 2) The systems perspective provides a way of defining the processes that establish a closed system (Fig. 10b) thus defining feedback in terms of system output. Most analysis of feedback concentrates on the global-mean climate system and global-mean surface temperature defining cloud feedbacks as those processes that connect changes in cloud properties to changes in global-mean temperature. There is, however, no theoretical basis to define feedbacks this way nor any compelling empirical evidence to do so. This is a point underscored by a number of studies that suggest this time–space-averaged perspective distorts the actual effects of nonlinear feedbacks that vary in space and time.
- 3) Most analyses of feedbacks assume that they operate independently of one another. A number of studies, however, suggest that some feedbacks couple together in a way that usually involves clouds. One of the major omissions in mapping from a complex system defined on the intrinsic time and space scales at which the cloud feedback processes take place to its steady-state global mean analog is the loss of the influence of the large-scale atmospheric circulation on clouds. It is argued throughout this paper that one step toward unravelling the complex nature of cloud feedback lies ultimately in understanding such influences. Diagnostic studies that touch on this topic were described in section 9.
- 4) The systems approach is also particularly cumbersome when applied to GCM experiments. The analysis in principle requires turning off processes individually to define the open system response, the estimate of which has led to much confusion in interpreting feedback in models. Comparisons of feedback diagnostics applied to the same GCM experiments but derived using different analysis methods with different assumptions about the nature of the system, each rooted to the systems framework approach, produce estimates of feedbacks that not only vary in strength but also in sign. Thus we are led to conclude that the diagnostic tools currently in use by the climate community to study feedback, at

least as implemented, are problematic and immature and generally cannot be verified using observations. Clearly alternative methods for feedback analysis need to be encouraged, like the approaches implied in the reference to the system identification problem or implied in the studies of Aires and Rossow (2003), Lynch et al. (2001), Monahan (2000), and others. Any new approach, however, will have little value if not explicitly tied to observations and the challenge is to develop more integrated methods that combine model sensitivity studies to observations.

With these comments in mind, we now revisit other themes of this paper as posed in relation to the commentary of Chamberlain.

- *The dependence of clouds on system output:* This is the most critical yet least understood aspect of the cloud feedback problem. Although the cloud processes that influence the radiation budget, in principle, are numerous and occur over a vast range of scales, the dominant scale of variability of cloudiness is the synoptic scale (e.g., Rossow and Cairns 1995) and the dynamics of the atmosphere on this scale, to first order, establish the relation between cloudiness and temperature, the latter being the usual measure of system output. The processes that connect the general circulation of the atmosphere to the formation and evolution of the large cloud systems associated with weather systems, and the latent and radiative heating distributions organized on this larger scale, establish the most rudimentary aspects of the feedback cycle (Fig. 3).
- *The “absorbent” nature of condensed water:* Although the effects of clouds on the distribution of absorbed radiation of the planet can be inferred from TOA ERB measurements, partitioning these effects between the atmosphere and surface is not immediately available from satellite observations. Cloud feedbacks care about how this absorbed energy is partitioned within the column and this depends on the amount of condensed water, and other factors. For example, the amount of sunlight reflected from clouds and thus absorbed at the surface, to first order, is influenced by the total water path. The amount of heating within the atmosphere, on the other hand, is dictated by the vertical distribution of cloud water (e.g., Slingo and Slingo 1988; Stephens et al. 1994). Many cloud feedback studies have ignored the atmospheric heating by clouds and the links to dynamics that this heating provides, focusing on the effects of clouds on the TOA or surface energy budget. Yet CMIP data analysis demonstrated that perturbations to the atmospheric radiation budget induced by cloud changes dictate the eventual response of the global-mean hydrological cycle of the climate model to climate forcing. Since clouds are a principal modulator of the atmospheric radiative heating, cloud feedbacks

are likely to control the bulk precipitation efficiency and associated responses of the planet's hydrological cycle to climate radiative forcings.

- *Cloud properties of relevance to feedback:* While many cloud properties exert important influences on the energy budget of the planet, not all of these properties are necessarily relevant to cloud feedbacks. Only those properties that depend on system output, by definition of feedback (e.g., section 6), define relevant feedback processes. Early feedback studies in the 1970s focused on cloud amount and cloud-top temperature, and these were followed by studies that examined feedbacks associated with optical properties (optical depth) through the connection to ice and water contents and the relation of these contents to (surface) temperature. We learn from the cloud LWP example that isolating the temperature dependence of these parameters, for example, is complicated because of the indirect influence of most of these parameters on airmass properties governed by the atmospheric circulation. At this time, these influences on cloud water and ice contents are neither well observed nor well understood.

Progress in understanding the cloud feedback problem has been slow for the reasons discussed. It has been argued that, in view of the complex nature of the climate system and the cumbersome problems encountered in diagnosing feedbacks, understanding cloud feedback will be gleaned neither from observations nor proved from simple theoretical argument alone. The blueprint for progress must follow a more arduous path that requires carefully orchestrated and systematic combination of model and observations documenting model improvements. Models provide the tool for diagnosing processes and quantifying feedbacks while observations provide the essential test of the model's credibility in representing these processes.

Although there are many aspects of the cloud feedback problem that are not well understood today, there are, however, reasons to expect progress in the coming decade.

- *Improved global-scale experimental data:* As previously mentioned, clouds are currently predicted in the most sophisticated cloud-resolving models (CRMs), NWP models, and climate models in terms of 3D distributions of cloud water and ice. Predictions of these fields cannot be validated in detail at the present time thereby thwarting model assessment and ultimate improvement. The availability of global-scale data on precipitation, albeit confined to the global Tropics (TRMM; Kummerow et al. 2000), as well as the near-future availability of global cloud water and ice information from CloudSat (Stephens et al. 2002), especially when combined with existing information from programs like ISCCP, provides the much needed datasets for evaluating cloud parameterization schemes and effects of clouds on the at-

mospheric radiation budget and water cycle under a variety of weather and climate regimes.

- *Improving the representation of clouds in models:* CRMs have evolved as one of the main tools for studying the links between key processes pertinent to cloud-related feedbacks. As such, these models may be viewed as an essential tool for articulating the underlying theories of cloud feedbacks, being adopted more widely in a variety of cloud and precipitation research activities. CRMs are also being used in experimental ways as an explicit form of cloud parameterization, thereby overcoming, in principle, the problematic separation between resolved cloudiness and unresolved convection. Despite the improvements of CRMs and their more widespread use, evaluation of CRMs is far from extensive, being limited to a few test cases from a limited number of field campaigns. The new global observations mentioned above will greatly advance CRM evaluations. However, the cloud evolution predicted by these models is sensitive to initial conditions (including the large-scale forcing that drives them). This sensitivity is problematic given that the source of this forcing usually derives from the analysis of large-scale operational models, the cloudiness from which is often suspect. Therefore progress in CRMs has to be intimately tied to progress in NWP global models. Mutual improvements, in turn, can be expected to lead not only to better cloud prediction schemes in global models but also can be expected to promote new assimilation methods applied to CRMs and eventually a more penetrating way of testing and improving models with observations.
- *Improving methods for testing models:* The recent years have witnessed the introduction of advanced diagnostic methods for evaluating cloud prediction in global models such as in the examples described in section 9. With increasing computational power expected in the coming years and the higher spatial resolution expected of these global models, continued improvements in the representation of smaller-scale cloud processes with the subsequent improvement in predictions of cloud properties is anticipated. Thus with improved resolution and expected improved global observations mentioned previously, more probing/testing of model parameterizations beyond that of section 9 can be expected. Presumably better parameterization methods will result, leading to better cloud predictions. Better cloud predictions, in turn, offer more capable assimilation methods, eventually expanding the use of existing and archived observational data, such as the archived but unused cloudy-sky radiance data derived from operational analysis.
- *The expanding role of NWP and data assimilation:* As previously emphasized, clouds tend to be organized into large-scale weather systems, or cloud systems, shaped by the global-scale atmospheric circulation.

Since the processes that govern cloud evolution are primarily a manifestation of the weather systems that form the vast cloud masses that dominate the energy balance of the planet, a fruitful strategy toward this goal presumably must embrace the study of weather including use of numerical weather prediction and related assimilation activities, in addition to the ongoing use of climate models. These studies should consider the problem of cloud evolution over a range of time scales including prediction of the diurnal cycle of clouds and precipitation. The combination of weather prediction models, complete with the extensive assimilation of global meteorological and satellite data and routine analysis to verify the forecasts, provides our most extensive and tested knowledge of the circulation of the atmosphere. However, progress in assimilation requires careful model evaluation and definition of model errors, which in turn, must rely on more extensive evaluation of parameterizations in models. Thus model development, evaluation, and assimilation remain highly coupled activities. These NWP-related activities, however, while necessary should not be considered at the exclusion of other cloud-climate research. Feedback analyses in the NWP context will not necessarily test subtle changes in processes that evolve on longer time scales associated with, for example, decadal changes in cloudiness and the radiation budget.

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