Climate Modeling

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Understanding and predicting climate change have recently acquired a sense of urgency with the advent of serious climate-related food shortages and with the realization that human activities may have an influence on climate. Unfortunately, there is no comprehensive theory of climate to explain its variability, nor are there physical models that can adequately simulate the climate system. Hence there is a serious need for the development of quantitative mathematical models of climate, first to derive better understanding and ultimately to allow some degree of predictive capability. Many modeling approaches are available, ranging from simple one-dimensional representation of the vertical radiative processes in the atmosphere up to very complex mathematical systems that describe the three-dimensional behavior of the circulation of the atmosphere and ocean along with the chemical and thermodynamical processes that control the hydrological cycle and the existence of sea ice. Since the simpler models isolate the important physical processes that determine the climate, we weigh our discussion more heavily toward these but also discuss many aspects of the more complex models. We try to stress the fundamental physical basis of each kind of model and its contribution to the understanding of the simultaneous interactions (or feedbacks) of the many coupled processes that determine the climate. One difficulty in predicting climate change is the large degree of cancellation that occurs between many of the climatic feedback mechanisms. Considerable discussion is included of the applications of theories of baroclinic disturbances to the parameterization of eddy heat fluxes. Also stressed are basic energy balance approaches to the calculation of the surface temperature. Finally, we give our view of the future of climate modeling and argue that a flexible plan that vigorously pursues many avenues of approach is preferable. Although this review is often critical, we hope that it serves (1) to illustrate the multidisciplinary nature of climate simulation modeling and (2) to clarify the relative importance of individual contributions to the broader problems of understanding and (ultimately) predicting climate change.

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A. INTRODUCTION

1. *Climate Modeling: What for?*

The climate is not constant. It has changed in the past, is changing now, and no doubt will change in the future. Furthermore, climatic variation important to human welfare need not be as spectacular as the last ice age—when (20,000 years ago) kilometer-high sheets of ice covered parts of the northern hemisphere as far south as 40° latitude. Since the retreat of the ice sheets 10,000 years ago there have been important climatic variations (for a recent examination of climatic variability see the paper by *Lamb* [1969]). For example, about 1000 years ago a relatively mild period permitted Viking exploration of the North Atlantic region. Several hundred years later, at the onset of the 'little ice age,' the Norse colonies in Greenland were wiped out, and the historical chronicles tell of long harsh winters that brought suffering and deprivation to Europe. Milder climate has returned in modern times, but the optimum (warmest) condition occurred in the 1940's, and since then, there has been a fairly rapid cooling in the high latitudes of the northern hemisphere. The relatively benign climate that we have taken for granted in the latter half of the twentieth century is not characteristic of all periods since the retreat of the ice age.

Climatic fluctuations are not new. In Biblical times (Genesis 41) Joseph warned Pharaoh: 'Behold, there come seven years of great plenty throughout all the land of Egypt. And there shall arise after them seven years of famine; and all the plenty shall be forgotten in the land of Egypt; and the famine shall consume the land.'

What is new is that the number of people affected by the possibility of starvation has increased. Now we talk of millions of people rather than the thousands of Biblical times.

Climate-induced famine is even more serious today. Already, 6 consecutive years of drought have ravaged large populations in parts of central Africa known as the 'Sahel.' The drought has left millions of people near starvation, and the number of deaths approaches hundreds of thousands. There are hints that the situation may yet worsen [e.g., *Winstanley*, 1973; *Bryson*, 1973]. At present, world population being high and food reserves being the lowest since World War II, ' "one big crop failure and it could be every country for itself." For most countries in the hunger belt that would mean mass starvation' [*Time*, 1974, p. 50]. Thus the understanding of natural climatic variability has taken on new urgency in our times [*Schneider*, 1974*a*].

But now there may be more than natural forces to be considered. The ability of humankind to increase in numbers and the material standards of life have depended upon the sustained application of technology.

Unfortunately, much of our technological progress has been accompanied by pollution as a disagreeable by-product. Our concern here is with the kinds of pollution that may in turn affect the climatic environment. One type of such pollution is the discharge of waste chemicals sufficient to alter the natural composition and radiative properties of the atmosphere. Another type of pollution involves changes to the radiative and water-bearing characteristics of the earth's surface; these changes originate from land-use practices that in turn modify the energy budget of the earth-atmosphere system. A third type is simply the release to the environment of thermal energy, a pollutant that now seems destined to increase more rapidly than the others, not only because per capita energy consumption increases with a society's standard of living but also because the means for abolishing other pollutants are almost invariably energy intensive [*Kellogg*, 1974*c*].

How is all this relevant to climate modeling? Of course, the climate of the earth has undergone many rather drastic changes in the past, and future changes appear likely [*Kukla et al.*, 1972]. The problem is not whether our climate will change but rather in what direction and from what causes it will be changed. There are now, however, other factors than there were in the past to be weighed in any speculation about future climate changes, for man is altering both the face of the earth and the composition of the atmosphere and is releasing energy into the global environment on such a large scale that his influence may no longer be dismissed. Mankind may have already caused climate changes, although no one can yet be sure to what extent [*Study of Man's Impact on Climate*, 1971].

Faced with the prospect that human effects on the environment may grow to the magnitude of natural forces and aware that severe climatic shifts (e.g., glacial periods) have been documented for the recent past, we would be prudent to consider urgently possible future effects. We need to know what kind and amount of environmental stress derived from man's activities might be tolerated by the climate system before that system would respond significantly. Such a question requires not a prediction of the detailed evolution of the resultant final state but merely a determination of the relative importance of the numerous interacting physical processes in the land-ocean-atmosphere-cryosphere (snow and ice) system. Equipped with this knowledge, we would be able to ascertain the sensitivity of the climate system to human perturbations (or nature's, e.g., volcanic eruptions) even without being able to predict the exact magnitude and course of the system's response. Our ultimate goal is climatic prediction, but it appears that we are not yet able even to speculate on the feasibility of long-term climatic predictability. However, recent developments spur the hope that more limited objectives (prediction of shorter-term annual or seasonal climatic anomalies and knowledge of climatic sensitivity on longer time scales) may now be within our grasp, and therein lies the indispensable immediate role of climate models.

Given the desire to understand the relative importance of various physical processes in the climate system, we now look for tools with which to build up a theory of climate and climate change. Since the earth-atmosphere system cannot be reproduced satisfactorily in the physical laboratory, we must augment empirical data with mathematical studies, i.e., climate models.

Our upcoming review of many of the existing climate-modeling efforts may appear to some readers to be overly critical and too often pessimistic in tone. Although we do of course feel that there is enormous room for improvements, we do not want to be interpreted as suggesting that these pioneering efforts have been unimportant to climate theory. Obviously, one must learn to crawl before learning to walk, and we hope that we have not given the impression that the imperfect models developed to date have been valueless in the building of a quantitative theory of climate.

Finally, we have largely confined our citations to current work. The reference list is extensive but by no means exhaustive even for the current literature, and apologies are given for those relevant references that we have inadvertently omitted. More classical works may be found through reference to our citations.

2. *Definition of 'Climate'*

To formulate mathematical approaches to a theory of climate, we must first define 'climate' as precisely as possible. We generally think of climate in reference to the average behavior of the land-ocean-atmosphere-cryosphere system over relatively long periods rather than to the detailed daily fluctuations that we call 'weather.' (For a general discussion of the collection and use of climatic data see chapter 1 of *Landsberg* [1967].) One problem of definition is to distinguish clearly between weather and climate. On the basis of recent conclusions about the predictability of weather [*Lorenz*, 1973] we could set the time scale division somewhat arbitrarily at 3 weeks. It is not practical, however, to define climate only in terms of 3-week time averages, for these would retain a large and unpredictable 'noise' component from the weather fluctuations. Instead, we might distinguish between what could be called an 'internal' system (e.g., the atmosphere) that is characterized by relatively rapid fluctuations and an 'external' system that provides relatively slowly changing external influences on the internal system. Climate could then be defined in terms of averages over a hypothetical ensemble of internal states that is nearly in equilibrium with the slowly changing external influences. By the terms of this definition, climate changes as the external conditions change. Whether changing influences have in fact produced significantly different internal states must necessarily in practice be decided on the basis of a limited sample. The sampling question is discussed further in section A5. (To help clarify what is meant here by noise, it is useful to repeat a comment by J. Smagorinsky (private communication, 1974): 'I certainly know that by noise you do not just mean an annoying high frequency variability trivially superimposed on the so-called signal. Although it may look like this statistically, it certainly isn't so mechanistically. The energy transformations connected with baroclinic instability and index cycles are essential interactive mechanisms for longer term variations. It seems to me they do not connote the usual meaning of noise.')

The division between such internal and such external systems is itself difficult to define precisely. A dry land surface with its rapidly changing temperature should certainly be considered part of the internal system, and most of the oceans could be regarded as part of the external system, but we are less certain about soil moisture, fluctuating snow cover, or ocean surface layers. However, in any numerical model used specifically for short-range weather prediction a distinction is always made between variables of the internal system as computed by the model and variables of the external system whose values may be adequately specified with no dependence on the internal state (e.g., the numerical models described by *Kasahara and Washington* [1971], *Manabe et al.* [1965], and *Somerville et al.* [1974]).

It is clear that as the time scale of climate change is expanded, the internal system will ultimately include all parts of the oceans, snow and ice fields, and possibly even part of the biosphere as well as the atmosphere; the external system will then reduce to those conditions truly uncoupled from the internal system: for example, the ocean and land topography and the distribution of incoming solar radiation.

This approach to defining climate (C. E. Leith, unpublished manuscript, 1973) is useful for clarifying the distinction between internal and external systems. However, it is simpler, especially in referring to observations, to retain the more conventional view of climate as a time-mean state (as described by *Landsberg* [1967]). In practice, time means are determined over a given interval of time, and 'climate change' refers to changes of state on a time scale greater than that of the given averaging time. Fluctuations on a shorter time scale are then regarded as noise. As is true in Leith's definition, it is desirable, at least conceptually, to remove the noise component (owing to shorter time scale processes) from an average over a given interval. That is, high-frequency noise over a period of a few days, for example, will leave a residual level of noise in a longer-term average over a given interval of, say, 30 days. Different 30-day averages of a time series with no long-term trends will still have slightly different means, because 30 days is a finite interval that contains a finite number of high-frequency fluctuations. In principle, the residual level of noise associated with different 30-day averages can be removed by ensemble averaging of an infinite number of 30-day means. In practice, of course, one cannot take an ensemble average; therefore the computing expense required to produce usable equilibrium statistics will depend upon how many realizations must be taken and averaged together in order to reduce the residual noise level sufficiently. This question is discussed in greater detail by *Leith* [1974] and in section A5.

When climate is viewed as a time-mean state, changes in the system on time scales that are long in comparison with a given interval may arise from either long-period external influences or other internal changes.

3. *Internal Versus External 'Causes' of Climate Change*

It has been popular to theorize that most climatic fluctuations are a result of changes in the external system, either naturally occurring or man induced, for example, (1) fluctuations in solar emission [e.g., *Roberts and Olsen*, 1973; *Dessler*, 1974; *Kellogg*, 1974*b*], (2) influence of variations in the earth's orbital parameters [e.g., *Vernekar*, 1972; *Mitchell*, 1972], (3) changes in atmospheric carbon dioxide [e.g., *Möller*, 1963; *Manabe and Wetherald*, 1967], (4) changes in atmospheric dust [e.g., *McCormick and Ludwig*, 1967; *Bryson*, 1968; *Charlson and Pilat*, 1969; *Lamb*, 1970; *Rasool and Schneider*, 1971], and (5) changes in the character of the land surface [*Study of Man's Impact on Climate*, 1971, chapter 7]. (Early suggestions as to climate change induced by these mechanisms or by mechanisms involving oceanic change are discussed by *Brooks* [1949].)

It is also possible for climate change to be inferred from natural time variations of the entire climate system without the presence of any external influence. (The internal system is now taken to include the entire interactive system of oceans, land, atmosphere, and cryosphere.) In this case the mechanism yielding this fluctuation is referred to as an internal cause. From the viewpoint of modeling, different time-average states representing internally generated climate regimes are derived either from prescribing the relevant internal cause or from specifying initial conditions that could lead a model integration to that average. If the internal cause is actually an internal degree of freedom of the model, then only the latter is applicable, and 'changing internal conditions' is synonymous with changing the model's 'initial conditions.'

Possible internal causes of climate change that have been discussed recently include (1) variations in ocean surface temperature patterns [e.g., *Namias*, 1972; *Bjerknes*, 1969], (2) decrease of North Atlantic salinity, leading to increased sea ice formation [*Weyl*, 1968], and (3) 'almost-intransitivity' of the land-ocean-atmosphere-cryosphere system [e.g., *Lorenz*, 1968, 1970].

With regard to the last cause of climate change Lorenz posed the question: Might not some scales of climate change be nothing more than natural fluctuations arising solely from the complex nonlinear interactions between land, oceans, atmosphere, and polar ice? These components of the climate system have the capacity to store or to release vast amounts of energy on time scales ranging from days to centuries. Fluctuations such as occur could be inherent in the complex natural climate system rather than a result of changes in the external environment (for example, concentrations of CO_2 or variations in solar input). Furthermore, it is possible that such natural fluctuations could occur on very long time scales (with respect to the observational data-gathering periods of man) and thus appear as distinct climatic states; in consequence, we might be tempted to look for external causes.

If the very long time averaged statistics of a climate system (or the set of equations that model the climate system) are independent of the initial state, the climate is said to be transitive. If no unique set of such 'equilibrium' climatic statistics exists, the climate is called intransitive. However, if the climate system (or model) is transitive for statistics taken over an infinite period yet contains shorter-period self-fluctuations that might be observed by man as climate changes indicative of an intransitive climate system, then this kind of climate system is called almost-intransitive by Lorenz. Therefore it is possible that almost-intransitivity could be interpreted as a cause of observed climate change. Thus we are faced with the possibility that for fixed external conditions, widely different climate states may be realized with finite interval time averages owing solely to intransitivity or almost-intransitivity. Without the assumption (as yet unfounded) of climatic transitivity we cannot speak of a particular climate as being associated with any given external state. Observed climate may change because of internal or external causes or some combination of both. In modeling, it is crucial for understanding a particular model to test the sensitivity of the equilibrium state of its statistics to variations of internal (e.g., initial) conditions. In connection with Leith's definition of climate (see section A2) an ensemble average is equal to an infinite time average only if the climate system is transitive. Another reason for retaining the traditional definition of climate as a time-mean state is that transitivity has yet to be proved for the climate system.

4. *Climatic Predictability*

A number of recent studies utilizing atmospheric data, numerical models, and turbulence theories have indicated that the motions of an atmosphere with fixed boundary and other external conditions are of limited predictability [e.g., *Lorenz*, 1973]. Even for a 'perfect' atmospheric forecasting model the inevitable uncertainties of determining the state of the atmosphere at any given time introduce errors at the beginning of a forecast that owing to the essentially nonlinear and unstable nature of the atmospheric dynamic system must grow and finally destroy any predictive knowledge of the true state of the atmosphere. These theoretical studies are more or less in agreement with each other and with present operational experience with using forecast models. (No such similar studies exist for the oceans, an important component of the climate system.)

An interesting analogy to the atmospheric predictability problem has been discussed informally by Lorenz. Consider a pinball machine in which a rolling ball strikes a cylindrical post (pin) and is reflected. The ball travels on a straight path until it encounters a second post, is reflected, travels a straight path, etc. Now, it is easy to visualize that if the ball strikes a piece of dust on the first path and is nudged slightly, the path following the first reflection will be slightly different from the non-dust-encounter case. It would probably not take more than a few strikes before the angle of reflection from a particular pin of the rolling path would be sufficiently different from the nondust case that the ball would entirely miss a pin that it had encountered in the first case; thus the subsequent path would bear no resemblance to the nondust case. It is clear that pinball predictability (i.e., which posts a particular ball will hit) is severely limited by the extreme sensitivity of the evolving trajectory to the exact initial velocity and direction of the ball. With a good knowledge of the initial trajectory it might be possible to predict four or five ball-to-post encounters with some confidence. If the initial trajectory is poorly known, this predictability will decrease to at most a few ball-to-post encounters. However, if the experiment is repeated many times, a probability distribution function describing the chances for a particular ball-to-post encounter can likely be found. Thus although it may be nearly impossible to predict the exact trajectory of an individual rolling pinball, it is quite probable that a statistical distribution of trajectories can be obtained that will be valid over a sufficiently large number of rolls.

Quite analogously, the instantaneous state of the atmosphere is not predictable much beyond 3 weeks, regardless of the accuracy of the (imperfect) specification of the initial state. (Of course, a poorly known initial atmospheric condition will reduce the predictability to well below 3 weeks.) However, as is true in the case of the ball-to-pin encounters, it may be possible to predict the statistical behavior of the atmosphere (i.e., spatial and temporal means, covariances, or higher moments of climatic state variables), given the external boundary conditions. For example, given the incoming distribution of solar radiation and the sea surface temperature distribution, it might be possible to predict the frequency and location of storm tracks across North America with more accuracy than could be obtained from statistical summaries of past storm tracks, even though the details of any particular track would be unpredictable.

It would seem, then, that the conclusions of atmospheric predictability studies rule out the possibility of any detailed long-range forecast for periods beyond the predictability limits. However, two simplifying assumptions have been introduced into these predictability studies, as pointed out by C. E. Leith (unpublished manuscript, 1973), and to the extent that they do not hold true for the real atmosphere, we again have some hope for long-range prediction.

The first of these assumptions, made at least in the turbulence theory approaches, is a sort of ergodic hypothesis that rules out the possibility that a growing error will temporarily level off at a value considerably below its ultimate limiting amplitude, defined as the point at which the predicted state becomes equivalent to a state chosen at random. This possibility will be realized if there exist separate regimes of internal behavior such that the atmosphere, having entered one regime, tends to remain there for extended periods of time. Studies dealing only with the growth of small errors completely avoid this issue. The evidence for persistent regimes seems unconvincing at present (see also the discussion of climatic transitivity in section A3), but if they do exist, positive forecast skill can be achieved simply by predicting that the present regime will continue; the existence of various regimes would be still more significant if the transitions from one regime to another were predictable.

The second assumption is that influences external to the atmosphere, such as incoming solar radiation or sea surface temperature, remain fixed. Influences external to the atmosphere obviously do change, and so long-range forecasting may be possible if we can not only forecast such changes but also determine the response of the atmosphere to such changes. One example might be the decrease of effective solar radiation caused by the presence of dust in the stratosphere. We may be able to predict the way in which such dust is slowly removed over a period of a year or two as a basis for a long-range atmospheric forecast. Volcanic eruptions are not now predictable, but the consequences of such eruptions may be. Another commonly considered example is the influence of anomalies in sea surface temperature. Such anomalies persist for months and could serve as predictors of atmospheric anomalies if the atmospheric response were known.

5. *Climatic Signal, Noise, and Equilibrium*

We have defined climate in terms of the average behavior of the atmosphere and have pointed out that knowledge of changes in conditions external to the atmosphere may be used as a basis for climatic forecasting. Furthermore, knowledge of distinct climatic regimes may also be useful for forecasting even if such external conditions remain unchanged. It is first necessary, however, to know what changes in atmospheric statistics are implied by changes in initial or external conditions.

This information can be derived by numerical experimentation only if it is not obscured by the unpredictable day-to-day variations of the model-generated atmosphere. We must answer the question as to whether the atmosphere is sensitive to a given change for a given averaging procedure. Sensitivity is defined as the realization of an induced, statistically significant signal in the presence of the noise of internal day-to-day weather variations.

The question of the sensitivity of atmospheric general circulation models is now being addressed. *Washington* [1972] and *Warshaw and Rapp* [1973] have given insight into this problem by comparing the magnitude of their model's response to a random perturbation. More recently, *Schneider and Washington* [1973] have tested the response of the National Center for Atmospheric Research (NCAR) six-layer general circulation model (GCM) to positive and negative global anomalies of 2°K in the sea surface temperature boundary condition in order to study the possible role of cloudiness as a climatic feedback mechanism. The model predicted (for a 30-day time average and a global spatial average) a change of -2.74% and $+1.53\%$ in low cloudiness for a positive and negative 2°K anomaly, respectively. (A

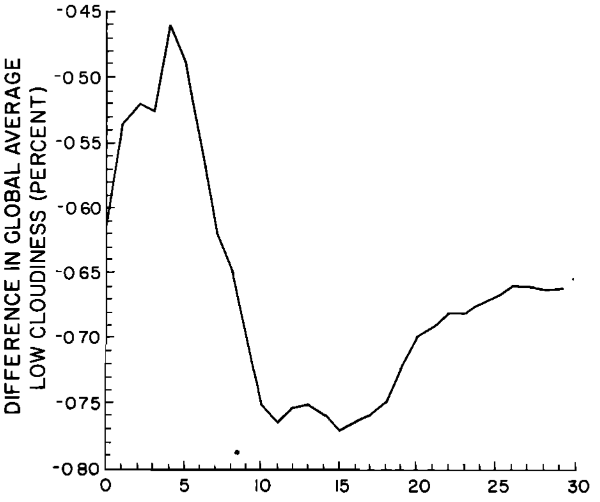

Fig. 1. Global average difference in low clouds between randomly perturbed and control runs with the NCAR GCM as a function of the number of model days of time averaging. This is a measure of the 'noise level' in the model.

similar result was obtained around 1971 by S. I. Rasool and L. Umscheid (personal communication, 1971), using a two-layer version of the Mintz-Arakawa GCM.)

It may not be possible to establish easily whether such a result corresponds to physical reality (see sections B2*d* and B2*g*), but we can ask whether it is significant for the model. That is, would a small random perturbation in initial conditions be likely to cause a response equally as large as or larger than that of the anomaly experiment? If so, it would be difficult to attach any significance to the anomaly experiment, since it caused a response no larger than the model's noise level. Thus to test the significance of the 2°K anomaly experiment described above, *Chervin et al.* [1974], using the same equations and boundary conditions, ran a further experiment with a 1-m/s random perturbation in the initial zonal wind and compared it with the control experiment.

The results for low clouds are shown in Figure 1, in which the algebraic difference between random and control experiment cloudiness is plotted as a function of the number of days of time averaging. Figure 1 suggests that the NCAR GCM, averaged over 30 days, would generate about 0.4–0.8% globally averaged changes in low cloudiness from nothing more than a random perturbation in initial zonal wind. In this case, then, the 2°K anomaly experiment produced changes in global cloudiness several times larger than the random experiment (−2.7% and 1.5%). Thus it is possible to conclude that the anomaly experiments produced a significant response in cloudiness. In other words, the 30-day mean model cloudiness is indeed sensitive to a 2°K change in sea surface temperature.

We might have expected the results of the cloudiness anomaly experiments to be significant, since the response to an anomaly of equal magnitude but opposite sign was comparable in magnitude and opposite in sign. Although the fraction of low cloudiness with and without a boundary perturbation was significantly different in this case, the zonal wind difference between anomaly and control experiments turned out in many places to be even smaller than the difference between the zonal wind in the random perturbation run and the zonal wind in the control run. Thus the noise level of the model in this dependent variable is apparently larger than the response of the model to the given anomaly in sea surface temperature, so that the zonal wind response to the anomaly, if any, is obscured (at least for 30-day means).

The noise level will depend of course upon the method of averaging, both in space and in time. For example, a 29-day time average of the random-minus-control temperature difference at 3 km for zonal spatial averaging (all external conditions remaining the same) is given in Figure 2, showing that model-generated noise levels from the equator to middle latitudes are substantially lower than those for latitudes in the vicinity of the poles. Sensitivity in this example is dependent on latitude.

It is also obviously important to decide how the accuracy of a climatic mean estimate depends upon the length of record used for computing averages. Figure 3 shows that the globally averaged root mean square (rms) difference in low clouds between the random-minus-control experiments described in the previous paragraphs is reduced by a factor of 2-3 by time averaging longer than 15 days. Similar results are obtained for other GCM variables.

An estimate of the reduction of noise with time averaging similar to that indicated by Figure 3 was obtained by *Leith* [1973]. Such an estimate is known from sampling theory to depend on the lagged time correlation of the time series, which for many meteorological variables can be reasonably well approximated by the exponential expression $R(\tau) = e^{-\nu|\tau|}$, where ν is a characteristic decay constant and τ is the lag time. This is of course the lagged correlation function for a first-order Markov process, sometimes called 'red noise.' If we define a correlation time

$$T_0 = \int_{-\infty}^{\infty} R(\tau)\, d\tau = \frac{2}{\nu}$$

which is about a week for the atmosphere, then for averaging times T much greater than T_0 Leith gives the asymptotic result that $\sigma_T/\sigma = (T_0/T)^{1/2}$, where σ is the standard deviation of the unaveraged time series and σ_T is that of the time average.

As an example we may consider the time series consisting of the maximum daily temperature at a mid-latitude location, where the standard deviation σ is of the order of 10°F (5.6°C). A 100-week time average is required in order to reduce the un-

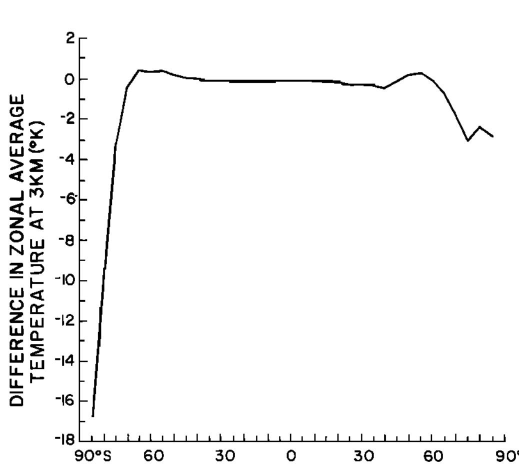

Fig. 2. Zonal average difference in 3-km temperature between randomly perturbed and control runs with the NCAR GCM as a function of latitude and for a 29-day time average, showing that noise level increases in high latitudes.

certainty of the mean to 1°F (0.6°C) by time averaging alone. Space averaging can provide additional reduction in error but only if the averaging is over regions with dimensions that are large in comparison with the space correlation length, which is typically about 2000 km. The observed increase in noise of longitudinal averages in the polar region is in part at least a consequence of the shorter space-averaging intervals there.

Thus the difficulty in detecting a small climate change by direct statistical methods may be so serious in some cases that it could call into question our ability to guard against just such small changes—which might be economically significant nonetheless. As an example, consider an 'undetectable' fractional change in the equator-to-pole temperature gradient. *Bryson* [1973] argues that the maximum poleward extent of the atmospheric subtropical high-pressure cells is directly proportional to the equator-to-pole temperature gradient. Furthermore, the northward penetration of the West African monsoon, which brings life-supporting rains to the south Saharan region known as the Sahel is also, according to Bryson, coupled to the position of the subtropical anticyclones and in turn to the equator-to-pole temperature gradients. Since a shift of 1° in the latitude of the Sahelian monsoon can, the argument goes, be triggered by a change of only a few tenths of a degree in the equator-to-pole temperature gradient, it is suggested that a directly undetectable climate change can lead to other changes significant for human welfare. (By 'direct' is meant a test on the primary climatic variable, i.e., temperature, in this case.) Carrying the example further, Bryson argues that an increase of a few tenths of a degree in the equator-to-pole temperature could be a consequence of increased atmospheric aerosol loading, among other factors.

Regardless of the validity of this example it serves the present purpose of illustrating that direct statistical tests of the detectability of climate changes may require supplementary inferences from either theoretical arguments or indirect statistical tests—which usually require knowledge of some cause and effect links between the direct and the indirect climatic variables.

B. INGREDIENTS OF A THEORY OF CLIMATE

1. *Physical Factors Affecting Climate*

A brief review of some of the basic physical factors that will have to be included in any quantitative theory of climate (following closely the paper by *Schneider and Kellogg* [1973]) is appropriate at this point. The fundamental determinants of the climate of the earth-atmosphere system are the input of solar radiation, the composition of the earth's atmosphere, and the earth's surface characteristics. Over a sufficiently long time the absorbed portion of the incoming solar energy must be balanced by the outgoing planetary infrared radiation. The temperature dependence of the latter determines to lowest order the mean temperature of the earth-atmosphere system. The optical properties of the atmosphere and the underlying surface determine the amount and location of the absorption of solar energy, the emission and absorption of infrared radiative energy, and consequently to a large extent the geographic distribution of heating of the atmosphere. The troposphere is essentially transparent to much of the solar radiation, which consequently is absorbed at the earth's surface and either warms the surface or evaporates water, energy subsequently being released into the atmosphere in the form

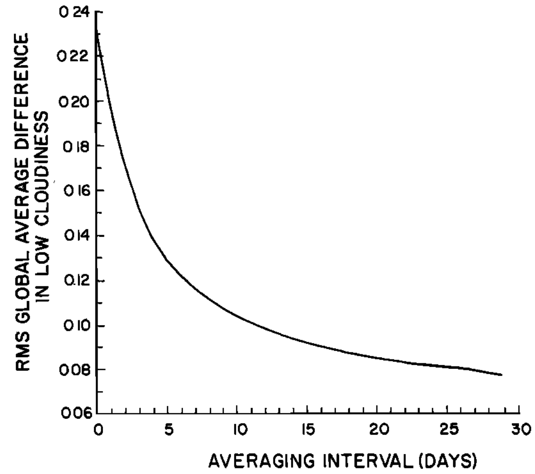

Fig. 3. Global average rms difference in low clouds between randomly perturbed and control runs with the NCAR GCM as a function of the number of model days of time averaging, showing reduction in noise level with time averaging.

of latent heat. This heating varies primarily with latitude but also has important longitudinal gradients as a consequence of the distribution of land and ocean surfaces. In the equatorial latitudes the absorbed solar energy generally exceeds the outgoing infrared energy, whereas in the polar latitudes the incoming absorbed solar energy is generally exceeded by the outgoing infrared energy. (It is not possible entirely to exclude significant time variations in the total energy output of the sun (the solar 'constant'), since its constancy has not been established to better than 0.5% [*Study of Man's Impact on Climate,* 1971].)

On the global average and over a long period of time a balance of planetary radiation is established, but this is not generally true locally in time and space. The rates of solar heating and infrared cooling are highly variable, not only horizontally over the globe but also throughout the vertical extent of the atmosphere. This unequal or differential heating of the globe, coupled with the rotation of the earth, is the ultimate driving force behind the winds and ocean currents. These winds and currents regulate the distribution of temperature, cloudiness, and precipitation over the globe by mutually adjusting with the radiative processes through transport of heat from areas of positive radiation balance to areas of negative balance. In this process the jet streams, trade winds, and eastward (westerly) winds of the middle latitudes, the westward (easterly) winds of the polar latitudes, and the migratory large-scale weather systems, or eddies, are all generated [*Lorenz,* 1967]. The circulation systems become more vigorous with increasing north-south temperature gradients applied at the lower boundary, so that large-scale transient eddies (storm systems), which transport additional heat poleward, provide 'negative feedback,' lessening the increase of the equator-to-pole temperature difference [*Thompson,* 1961].

The atmosphere conveys heat in two forms: 'sensible' and 'latent.' Transport of sensible heat involves the direct transport of warm air to a cold region. Latent heat is carried by the water vapor evaporated at the earth's surface. In the presence of suitable nuclei (particles) and saturation the water may condense into drops, thereby releasing the latent heat that was needed originally to change it from liquid to vapor. The process of evaporation, transport of water vapor, condensation, precipitation, and reevaporation (i.e., the hydrological cycle) is responsible for one fourth to one third of the net heat transported across the 30°N and 30°S latitude circles [e.g., *Budyko*, 1971]; sensible heat transport by the atmosphere accounts for another one fourth to one third of the total, and the oceans carry the remainder, between one half and one third of the total heat flowing poleward [e.g., *Vonder Haar and Oort*, 1973].

The energy balance of the earth-atmosphere system given as a function of latitude is illustrated by Figure 4 (derived from empirical data by *Sellers* [1969]), where R_e is the net zonal radiation balance, which (in equilibrium) is equal to the sum of the net divergence of horizontal energy fluxes associated with transport of atmospheric sensible heat ΔF_A, transport of oceanic sensible heat ΔF_O, and transport of latent heat of fusion by the atmosphere ΔF_q. (Other energy flux data have recently been discussed by *Newton* [1972].)

No summary of important climatic factors would be complete (at least if long-term climate changes are to be considered) without mention of the cryosphere, which includes the substantial areas of the earth's surface that are covered by ice and snow. The albedos of snow and ice are usually much higher, and the temperatures much colder than those of uncovered land or open ocean (see section C3*b*). Thus positive feedback between ice cover and temperature may be inferred: colder temperatures cause more ice and snow and thus a higher albedo with a consequent reduction in absorbed solar energy, which in turn implies yet colder temperatures.

We must keep in mind, however, that ice and snow with their high albedos are for the most part confined to limited regions of the earth, whereas globally, clouds reflect the greatest percentage of incoming solar energy. In general, the hydrological cycle, which includes the cycling of water in clouds and (in its solid phase) in snow and ice fields, looms as a major factor in determining mean surface temperatures, not only through its influence on snow, ice, and clouds but also through its control of surface vegetation and soil moisture. Since the hydrological processes are also tied to the motions of the system, it is clear that the radiation balance and the dynamics of the atmosphere-ocean-cryosphere system are tightly coupled through the hydrological cycle as well as by the direct dependence of radiation on temperature; therefore any valid quantitative theory of climate will ultimately have to treat the hydrological cycle in detail.

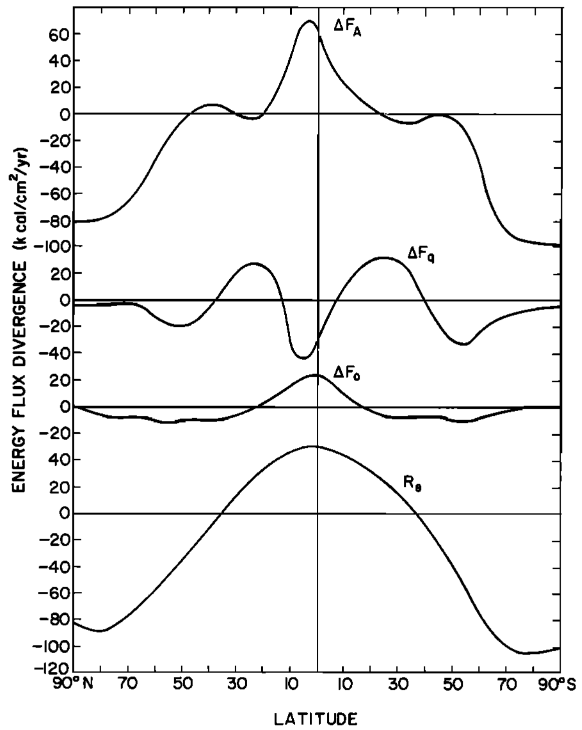

Fig. 4. Net horizontal flux divergence of atmospheric sensible heat ΔF_A, latent heat ΔF_q, and oceanic heat ΔF_O transport, which are balanced by the radiation balance of the earth-atmosphere system, R_e [after *Sellers*, 1969].

2. *Assorted Climatic Coupling Mechanisms*

Some possible climatic feedback mechanisms are listed here (mostly those of *Schneider and Kellogg* [1973]) in order to indicate various types of coupled processes that must ultimately be included in a 'realistic' climate model. Of course, models of varying complexity will contain the different mechanisms to varying degrees of realism. A more detailed account of the extent to which different types of models incorporate climatic feedback processes is presented in subsequent sections.

a. Temperature-radiation feedback. Thermal radiation depends on absolute temperature. An increase in the temperature of a substance results in an increase in the amount of thermal energy radiated from the substance (all other factors, such as chemical composition, remaining unchanged). The increased thermal radiation will act to restore temperature back to its equilibrium value. This process tends to limit or to stabilize the temperature response of a substance to changes in energy input. Therefore temperature and radiation couple to each other in a stabilizing fashion, i.e., in negative feedback.

b. Water vapor-greenhouse feedback. The atmosphere is believed to maintain a somewhat uniform distribution of relative humidity over a large range of lower atmospheric temperatures [*Möller*, 1963] even though the absolute amount of water vapor in the air varies strongly with atmospheric temperature. The absolute amount of water vapor in the atmosphere determines to a large extent the opacity of the lower atmosphere to infrared radiation. Thus increased atmospheric temperature at constant relative humidity leads to increased trapping of thermal radiation ('greenhouse effect'), which gives rise to further increases in temperature of the lower atmosphere. Therefore the coupling of temperature to the water vapor-greenhouse effect acts to destabilize, i.e., results in positive feedback.

c. Snow and ice cover albedo-temperature feedback. The high reflectivity of snow and ice as compared to that of water and land surfaces is a dominant factor in the climate of polar regions. However, the extent of the snow and ice cover of the earth's surface depends strongly upon surface temperature. Thus if lowering the planetary temperature would lead to a longer lasting and more extensive snow and ice cover, this would increase the planetary albedo, causing a decrease in the amount of solar energy absorbed by the earth-atmosphere system, and would thereby lower the temperature further. This positive feedback has caused many scientists [*Study of Man's Impact on Climate*, 1971, chapters 6 and 7] to consider the question of the permanency or stability of the polar ice caps, since departures from the present extent of ice coverage might be able to maintain themselves.

It would appear from the results of at least some models that changes of more than a few degrees in the planetary temperature or variations of more than a few percent in the energy input to the earth (see the discussion of energy balance climate models in later sections), if they are sustained over a sufficiently long period, could result in either a rapid expansion or a complete disappearance of the polar ice sheets as a result of the positive feedback between ice cover and temperature. However, this snow and ice cover albedo-temperature coupling must also be viewed in the light of other processes that might modify these conclusions. For example, hydrological processes should also be included in the feedback loop, since ice and snow are merely the solid phase of water. Thus the strong positive link between the extent of snow and ice cover and the local temperature assumed in the preceding discussion will be effective only insofar as there is an appropriate amount of precipitation, for example, to build up continental glaciers or snow cover on sea ice. *Kellogg* [1974*a*] has discussed other feedback processes occurring in polar latitudes; the present example illustrates the general complexity of the land-ocean-atmosphere system, with hydrological processes depending upon the atmosphere and upon global cloudiness. We now turn to some of these other interactions.

d. Cloudiness-surface temperature coupling. Most clouds are both excellent absorbers of infrared radiation and good reflectors of solar energy [e.g., *London and Sasamori,* 1971]. Thus changes in cloudiness of the order of even a few percent can have an effect on the radiation balance and hence on the surface temperature that is quite significant, as may be demonstrated with horizontally averaged models (as discussed in sections C2*a* and D1*a*). Because both solar energy absorption and planetary infrared emission decrease with increasing cloud cover amount, we must consider quantitatively the geographic distribution of cloud amounts, cloud heights, and cloud optical properties in order to calculate the net effect of changes in cloud parameters on the surface radiation balance.

In order to understand the details of cloud formation it is also necessary to refer to the cloud microphysics. Clouds would generally not form at all without the presence of suitable particles to nucleate cloud droplets. The development of droplets with a spectrum of sizes in a cloud depends on the nature of the available cloud (water and ice) condensation nuclei as well as the temperature, the relative humidity, and the motion field of the cloud. Droplets must grow to drop sizes of 100–200 μm or greater in order to be able to survive the fall to earth without complete evaporation. If the upward vertical motion is sufficiently weak, the rate of drop growth is sufficiently slow, or the air beneath the cloud is sufficiently dry, drops will evaporate below the cloud and hence will recycle their humidity to the air entering the cloud. Precipitation is usually generated in mid-latitude, layered cloud systems through the nucleation of ice particles, which can grow rapidly at the expense of water droplets. Changes in various climatic parameters, including the distribution of condensation nuclei, are likely to change the relative fraction of precipitating versus nonprecipitating clouds as well as the cloud droplet spectra. For these reasons, even in the absence of changes in precipitation, mean optical properties of clouds may change—with climatic consequences [*Pueschel et al.,* 1974].

The dependence of the above factors on temperature, and hence even the direction of possible climatic feedback between cloudiness and surface temperature, is not yet clear and remains to be determined through further observational and modeling studies that include the effects of microphysical, dynamic, hydrological, and radiative processes (see sections A5 and D1*a* for examples of attempts to model this feedback).

e. Radiative-dynamic coupling. Changes in the radiation balance initially caused, for example, by an increase in CO_2 will result in a redistribution of this heat by atmospheric motions, which may either offset or accelerate any climate changes linked initially to the original perturbation in radiation balance. The direction of any such coupling would have to be determined by specific model experiments using dynamic models of sufficient complexity to include all important feedbacks.

f. Ocean-atmosphere coupling. Although the intrinsic link between the oceans, the atmosphere, and the climate has been implied in preceding discussions, it is of sufficient importance to list this coupling separately. In addition to the obvious role of the oceans in providing water for the hydrological cycle the dynamic coupling of atmospheric winds and temperatures with ocean circulation and sea surface temperatures plays a major role in determining our climate.

The vast thermal capacity of the oceans limits the extremes of seasonal climate that would otherwise be experienced in the middle and polar latitudes were it not for the presence of the oceans. This thermal 'flywheel' effect of the oceans increases the response time of the surface temperature to changes in external energy input [*Manabe and Bryan,* 1969]. Thus the interactive role of the atmosphere and oceans in shaping the climate over the long term cannot be overstressed (see also the discussions of climatic intransitivity in sections A3 and A4 and ocean modeling in section C3*d*).

g. Lapse rate-surface temperature coupling. One extremely important coupling related to all those listed above is the link between surface temperature, humidity, and tropospheric lapse rate. The importance to climate theory of understanding this coupling is crucial. For example, computation of the surface temperature response to a given change in global albedo (from aerosols or clouds, for example) will depend on how the lapse rate is maintained.

Conventionally [e.g., *Manabe and Wetherald,* 1967], a convective adjustment that constrains the lapse rate to values less than either the dry or the moist adiabatic (or some average of these) lapse rate is used to determine the tropospheric temperature profile (as discussed further in section D1*b*). *Stone* [1973], on the other hand, has argued on the basis of a simplified climate model (discussed further in section D3*d*), which includes parameterizations for horizontal and vertical eddy heat fluxes but ignores moist convection and ice-albedo feedback, that the vertical heat transfer by large-scale eddies (i.e., $\langle w'T' \rangle$) may be largely responsible for maintaining the tropospheric lapse rate. This question, whether vertical heat transfer by moist convection or by large-scale eddies is most influential in determining the static stability, can only be resolved by considering models that contain adequate parameterizations of both. As a result of the strong negative feedback of vertical eddy heat fluxes Stone's model predicts that the mean static stability (i.e., the difference between the dry adiabatic temperature lapse rate and the actual rate) will change by only a few percent with a factor of 2 change in solar radiation, even though the ground temperature changes by as much as 50°–100°C.

This result is to be contrasted with a process that would maintain a moist convective lapse rate, where the static stability would be sensitive to surface temperature, a surface temperature change being amplified in the middle of the troposphere according to the temperature dependence of the saturation mixing ratio. One possible implication of this dependence has been discussed by *Kraus* [1973]. It can be represented schematically, as has been done in Figure 5, in which the moist adiabatic lapse rate is shown for two values of surface temperature. Numerical experiments currently being carried out with the GCM at NCAR by Schneider and Washington (discussed in section A5) show a positive feedback between cloud amount and surface temperature and tend to support the concept of a large sensitivity of tropospheric static stability to variations in solar constant through control of lapse rate by moist convection. However, it is difficult to determine whether these GCM's have underestimated $\langle w'T' \rangle$. The change in lapse rate with surface temperature is determined by the competing feedbacks between (1) surface temperature and $\langle w'T' \rangle$ or (2) surface temperature and moist convection. The lapse rate in turn largely determines the direction of the cloudiness-surface temperature feedback. Other aspects of this question are discussed in section D3*d*.

3. *Approach to Mathematical Modeling of the Climate*

The factors governing climate (solar radiation, ocean circulation, ice cover, winds, etc.), as discussed briefly in the previous sections, must all be related by the equations of the general climate theory. The principles of conservation of mass, momentum, and energy, taken together with the thermodynamical and chemical laws governing the change in material composition of the land, sea, and air, make up the fundamental theoretical basis for the theory of climate. This set of coupled nonlinear three-dimensional partial differential equations are to be solved subject to the external input of solar radiation and for a given initial state of the earth-atmosphere system.

Each variable is related to the others in such a way that changes in one invoke simultaneous variations in others that in turn have a feedback effect on the original variable, as discussed extensively in section B2 above.

The motions in the system take place on a wide range of time and space scales. Whereas features of the general circulation with a scale of 1000 km persist for days or longer, small-scale quasi-random motions in the atmosphere and the oceans may last only for seconds and occupy only a few meters of space but nevertheless can be of crucial importance in the turbulent transport of heat, momentum, and atmospheric or oceanic constituents at the air-surface interface. For practical applications these small-scale processes are treated statistically (if at all), their effects being related to average conditions over much longer periods and scales. The technique of relating the statistical effects of processes that cannot be computed in detail to those processes that are computed in detail is called 'parameterization' (short for parametric representation).

In order to solve the equations of a climate theory with the knowledge and tools available to us in the foreseeable future it is necessary to ignore the details of small-scale motions occurring on a continuous spectrum of scales and to treat the system at a discrete number of points in time and similarly use a 'grid' of points or some other form of discrete representation in space (such as Fourier modes). Then all motions occurring on scales smaller than the size resolved by the discrete spatial representation must be treated by parameterization or must be completely ignored. The technique of selecting appropriate spatial resolution and representation, time step size, and parameterization technique for use in the construction of approximate theories of climate is a large part of the art of 'modeling,' and the particular choice of these elements, along with the choice of physical and chemical factors, determines the model.

4. *Diagnostic Data for Model Validation*

Reproduction of the real atmosphere is clearly desired in our theoretical models, and so we must constantly inquire as to the behavior of the real atmosphere. Different classes of models, as discussed later, require different types of data for validation, depending upon the spatial and temporal scales dominant in the particular model. For example, longitudinally averaged energy balance models call for accurate data on energy transported across latitude circles by the mean meridional motions and eddy fluxes of sensible and latent heat in the atmosphere as well as energy transported by ocean currents. On the other hand, horizontally averaged vertical models need detailed information on vertical atmospheric heating rates, optically active gases, clouds and particles, optical characteristics of the earth's surface, and incoming and outgoing fluxes of both infrared and solar radiation.

For some models such as these, time averages of at least several years are probably most appropriate. However, validation of large-scale three-dimensional models of the earth-atmosphere system calls for a vast three-dimensional collection of data of basic fields with detailed time resolution: pressures, temperatures, humidity, winds, ocean salinity, ocean currents, cloudiness, and so forth. (The observational requirements for model improvement on time scales of a few weeks or less are extensively discussed by the *Global Atmospheric Research Program* [1973].) Also, derived statistics, such as outgoing radiation fluxes or eddy fluxes of momentum, energy, or moisture, should be compared between the atmosphere and the model.

Finally, valid large-scale models in turn can be used to provide statistics needed for the purpose of validating or constructing parameterizations for models of lesser complexity, such as dynamic models of the atmosphere that vary with height and latitude only.

It may prove useful for model validations to make use of

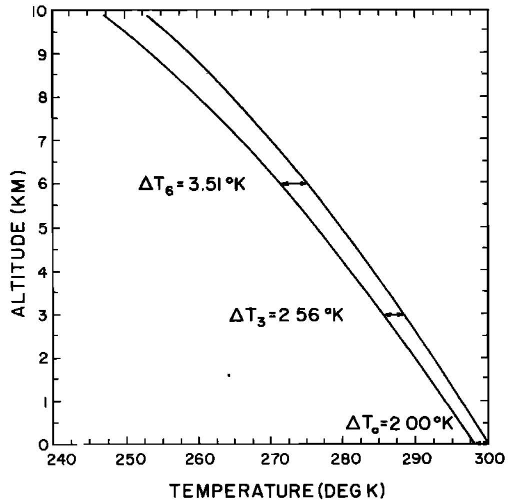

Fig. 5. Plot of moist adiabatic lapse rate versus temperature for two different surface temperatures, showing the decrease of moist lapse rate with increased surface temperature and thus the amplification of a surface temperature change with height.

the fact that the atmospheric circulation of the southern hemisphere is more zonal in character than the circulation of the northern hemisphere (due mainly to the hemispheric differences in continentality). A realistic model should be capable of simulating such interhemispheric differences. Papers providing much observational material on the southern hemisphere were recently collected by *Newton* [1972].

C. MODELING METHODOLOGY

1. *Classification of Climate Models*

It is helpful to attempt to categorize the different approaches that have been used for climate models, both to focus better on the objectives of the models and to indicate directions for future research. Many models investigate a single mechanism or a small number of simply coupled mechanisms, taking as being given (perhaps from observation but often by assumption) all the other components of climate. This type of model is directed toward understanding the dependence of the particular mechanism on the other parameters of the problem. Models of this sort can be referred to as 'mechanistic' models.

On the other hand, there are the 'simulation' models, which include as internal variables as many interacting physical processes as possible, the primary objective being to simulate observed phenomena. These models are often too complicated to allow their results to be interpreted unambiguously and usually require a great deal of analysis and computer time in order to provide much understanding of the individual mechanisms and their dependence on each other. It is thus difficult to trace cause and effect relationships or to interpret quantitatively which physical process or processes are responsible for causing a particular effect on the model-generated statistics, given a perturbation in one of the variables or boundary conditions. Thus the mechanistic models are used to suggest the relative importance of various terms in the simulation models and to indicate limited experimentation involving the parameter dependence of the simulation model.

Although it must be recognized that one ultimate objective of mathematical models of climate is to include jointly all the coupled feedback processes in a realistic fashion, it is unlikely that this can be done successfully without the understanding derived from simpler models of individual processes. The distinction between mechanistic and simulation models is not sufficiently sharply defined and alone does not do justice to the rich variety of modeling approaches that have been and will be used. Therefore we feel that a classification system based essentially on the number of geometric degrees of freedom contained in the model will be more useful, and so we summarize the hierarchy of models according to this scheme in the next section.

2. *Hierarchy of Climate Models*

The simplest models of climate are essentially one dimensional, dependence on other dimensions being simply parameterized or neglected. Thus we have, for example, one type of model intended to derive the vertical structure of temperature that assumes horizontal global averaging for the radiative and convective heat transfer processes. Other types of models parameterize the vertical dependences and determine variation with horizontal coordinates, sometimes only latitude or longitude.

The models with only horizontal degrees of freedom fall into two distinct classes depending on whether they have been developed to describe statistics of surface thermal balance ('energy balance' models) or of atmospheric dynamics ('barotropic statistical-dynamic' models). The energy balance models are extended by allowing for separate surface and atmospheric degrees of freedom, and the dynamic models by allowing two atmospheric layers ('two-layer baroclinic' model). At this stage the distinction between the two classes of models derives from the external specification of surface conditions in the case of statistical-dynamic models and from the simple parameterization of dynamic processes in the energy balance models. At the more advanced stage of model development, where sophisticated treatments of both atmosphere–surface energy budget coupling and the statistics of atmospheric dynamics are included, the two classes merge. Inclusion of detailed vertical structure then incorporates the processes described by vertical models. Thus a classification of climate models according to geometric degrees of freedom forms a pyramid, many kinds of simple models merging together as increasing complexity dictates inclusion of more internal degrees of freedom. The simplest models are largely mechanistic in their objectives, whereas the more sophisticated models are more intended for simulation. The pyramid of model hierarchy is briefly summarized in the following sections.

a. Horizontally averaged, one-dimensional vertical coordinate models. This type of climate model determines the various radiative fluxes as a function of the vertical coordinate and in some cases also gives the mean vertical temperature profile implied by the balance between net radiative flux and other vertical energy transfer processes. Such a model is most meaningfully employed in conjunction with horizontal averaging over the globe, although other cases can also be instructive. Globally averaged models that include the vertical distribution of atmospheric absorbers and scatterers can be used to study the relative magnitude of atmospheric radiative heating rates and the role that they play in determining the vertical structure of temperature. Besides radiative fluxes other processes, such as vertical convection, cloudiness, albedo, interaction between clouds and radiation, and water vapor distribution, are parameterized one dimensionally in these models. The impact of changes in CO_2 or aerosol distribution has been studied [e.g., *Manabe and Wetherald,* 1967; *Rasool and Schneider,* 1971]. Included are parameterized feedbacks between radiation, temperature, water vapor, and lapse rate. Many of the basic physical processes in the one-dimensional models are also parameterized in more elaborate models with equivalent (or lesser) vertical resolution. Hence the simple one-dimensional models by testing the sensitivity of radiative fluxes to perturbations in model parameters serve the dual role of developing parameterizations for more elaborate models and of providing first insights into the effects on climate of specific external changes, such as the addition of pollutants. It is not possible in one-dimensional vertical models to include adequately the contributions to global mean energy fluxes that arise from coupling horizontal variation of thermal fluxes with horizontal variation of other parameters, such as temperature, cloudiness, and ice and snow cover.

A deeper look into these models is given in section D1.

b. Horizontally varying energy balance models. The term energy balance models refers to those models that emphasize the calculation of surface temperature in terms of a balance between incoming solar and outgoing infrared radiation.

In the absence of an atmosphere or surface storage of energy and for a thermally black surface an energy balance model reduces to

$$\sigma T_p{}^4 = (1 - \alpha_p)Q_s \qquad (1)$$

where Q_s is the globally averaged solar energy flux, α_p the planetary albedo, σ the Stefan-Boltzmann constant, and T_p the planetary surface temperature. The radiative role of the atmosphere is incorporated into (1) by parameterizing atmospheric temperature in terms of model parameters. For example, a simple subset of the globally averaged models is derived by specifying the mean atmospheric lapse rate so that the temperature profile depends only on surface temperature. Upward and downward infrared radiative fluxes between surface and atmosphere, derived in terms of surface temperature and given parameters, can then be added to (1) to give a more realistic surface temperature calculation. Because of the simplicity with which vertical dependences are incorporated into such a model it is relatively easy to generalize the model to allow for horizontal variations. With such extension it is, however, necessary to parameterize the horizontal redistribution of energy by atmospheric and oceanic transport. One such approach is that of *Stone* [1973], who simply describes horizontal and vertical variation in terms of global mean latitudinal and vertical temperature gradients and energy fluxes and thus has derived a climate model that is essentially zero dimensional, though allowing simply for coupling of three-dimensional dynamic processes with the mean thermal field (discussed in more depth in section D3).

Models with more mathematical degrees of freedom are now described.

1. Energy balance models varying as a function of latitude only have been useful for simply examining the coupling between latitudinal variation of temperature and temperature-dependent albedo. Variable albedo has been introduced by parameterization of snow and ice cover in terms of zonal mean temperatures. The poleward transport of energy by oceans and atmosphere is parameterized in terms of surface temperatures. Energy balance models are sometimes referred to as 'semiempirical' models [e.g., *Study of Man's Impact on Climate,* 1971], since the temperature dependence of albedo and horizontal energy transport as well as infrared fluxes has generally been derived from empirical relationships.

Budyko [1969] thus simply related horizontal energy fluxes to the difference between zonal mean and planetary mean temperatures, whereas *Sellers* [1969] has attempted to relate the fluxes to horizontal thermal mixing by large-scale eddies and to a mean meridional wind. The extreme sensitivity of these models to changes in incoming radiation resulting from the positive temperature-albedo feedback has encouraged further study and generalization. Integration of time-dependent versions [*Schneider and Gal-Chen,* 1973] shows them to be quite stable to perturbations in internal (i.e., initial) conditions, as is discussed further in section D2*a*. Obviously needed areas of improvement include parameterization of albedo-temperature coupling, horizontal energy transport, and cloudiness distribution.

Because these models have shown such drastic results, their structure is discussed in further detail in section D2.

2. Models dependent upon both latitude and longitude are described next. The longitudinal variations of climate from continental to oceanic regimes are in many places nearly as pronounced as the latitudinal variations. A reasonable dynamic treatment of longitudinally varying atmospheric structure and transport is, however, more difficult than a treatment of latitudinal variations. This difficulty is avoided if the longitudinal variation of atmospheric parameters, such as temperature, humidity, and cloudiness, is specified. *Saltzman* [1967] has so modeled the longitudinal variation of surface temperature. The dominant factor determining surface temperatures over oceans in his calculation appears to be the upward mixing from the base of the seasonal thermocline, whose temperature is specified. Longitudinal variations of humidity, cloudiness, and convection at the surface (depending on advection of cold air near the surface) are largely responsible for forcing the variations of continental surface temperature.

Another approach to modeling longitudinal climatic variation arising from land-sea contrast is that of *Sellers* [1973], who extended his previously discussed zonal model by distinguishing separately between a land and an ocean temperature at each latitude. Essentially, he assumes only two grid points in longitude but allows for the relative areas of land and ocean around each latitude circle. His calculation of motions, as is true in his previous model, is parameterized in terms of surface temperatures. The complex dynamic processes redistributing thermal energy in longitude are replaced by simple zonal flow advection between land and ocean areas by using longitudinally averaged motions, as described later in section D2*b*.

A third energy balance modeling approach advocated by *Adem* [1970*a*, *b*] assumes that latitudinal and longitudinal atmospheric and oceanic thermal transports may be parameterized largely in terms of horizontal 'Austausch' coefficients. The most recent versions of his model allow for transport by observed mean winds and by wind-driven ocean currents.

Adem's models have been developed largely for attempts at forecasting climatic variations on a time scale of up to several months, and so surface albedos are prescribed rather than calculated. It would appear that the predictive capability of these models would depend largely on their sensitivity to initial and predicted anomalies in sea surface temperature. *Namias* [1972] has discussed the observational basis for relationships between anomalies of sea surface temperature and the longitudinal distribution of atmospheric pressure. He argues that sea surface temperature appears to change primarily through advection by wind-driven currents.

c. Zonally symmetric dynamic models of the atmosphere. The objective of this class of models is to derive the zonal structure of the atmosphere from an observationally prescribed distribution of thermal and momentum sources. The various models of this type are distinguished by which of these sources are given and which are parameterized in terms of the internal (zonal mean) variables of the system. The prototype for these models is the study by *Eliassen* [1952], who assumed all source terms to be prescribed. These sources drive a meridional circulation that redistributes heat and momentum in such a fashion that the zonal wind, pressure, and temperature fields evolve while they maintain hydrostatic and geostrophic balance. Steady state solutions are possible with Eliassen's approach only if the prescribed momentum sources are balanced by Coriolis torques due to mean meridional circulation, so that there is no net zonal wind tendency, and if prescribed heat sources are balanced by adiabatic cooling due to mean vertical motion, so that there is no net zonal temperature tendency.

External momentum and thermal sources can be prescribed independently, however, if there are also internal sources, depending on the variables of the model. One such approach is to attempt to parameterize the zonal mean eddy fluxes of heat and momentum in terms of the zonal structure. *Smagorinsky* [1964] has related eddy momentum flux to heat flux, assuming a two-layer model and a simple baroclinic disturbance. He then derived a differential equation for the eddy heat flux required by the observed distribution of thermal sources and the assumption of no net zonal temperature tendency. A vanishing of zonal acceleration was achieved by balancing Coriolis torque with momentum convergence in the upper layer and with surface friction in the bottom layer.

Another approach recognizes that part of the net thermal source, especially that involving infrared radiative emission, must be sensitive to zonal mean temperature, whereas part of the net zonal mean momentum source should be 'frictionlike,' that is, sensitive to the zonal mean winds. If the temperature-dependent part of the heating and the frictionlike part of the momentum source can be parameterized, the resulting system of equations provides a description of annual mean or slow seasonal variations of not only the mean meridional circulation but also the mean zonal winds and temperatures. This approach has been used by *Dickinson* [1971*a*, *b*] to discuss the relative roles of zonal mean eddy momentum transport and heating by the tropical rain belt in determining zonal winds, temperatures, and meridional circulation from the tropics to mid-latitudes.

In view of the great importance of heating from zonal mean rainfall, especially in the tropical rain belt, for zonally symmetric dynamics, one obvious further constraint to relax is the external prescription of latent heat release. For example, *Pike* [1971] in an elaborate primitive equation model for zonally symmetric dynamics in the tropics has parameterized the release of latent heat in terms of a hydrological cycle depending on mean meridional circulation, temperature, surface zonal wind, and sea surface temperature. A separate dynamic zonal ocean model responds to zonal wind stress and calculates the sea surface temperature, and so the feedback between atmosphere and ocean is investigated.

Other recent zonally symmetric dynamic models [e.g., *Saltzman and Vernekar,* 1971, 1972; *Kurihara,* 1970, 1973; *Wiin-Nielsen,* 1970, 1972] have included certain aspects of the surface energy budget (but have excluded a cryosphere calculation) and have attempted with varying degrees of elaboration to improve the parameterization of eddy heat and momentum fluxes. The question of the appropriate procedures for parameterizing eddy fluxes is crucial to the development of a self-consistent zonally symmetric model and so is given special consideration in section D3.

d. Atmospheric dynamic models emphasizing longitudinal variation. Longitudinal variations of mean atmospheric pressure patterns are strongly correlated with the variations of mean rainfall, cloudiness, cyclogenesis, and other such weather elements. Indeed, nearly a century ago Teisserenc de Bort pointed out the existence and importance of large-scale pressure patterns, which he referred to as 'centers of action.' With the availability of upper air soundings it became evident that the surface pressure patterns were closely related to patterns at the midtroposphere. *Rossby et al.* [1939] interpreted these disturbances in terms of a series of stationary waves generated by some combination of thermal and topographic sources and developing 'in somewhat the same fashion as the standing waves which are sometimes observed in clouds on the lee side of a mountain ridge.' Rossby suggested that the basic dynamics involved the balance between redistribution of planetary vorticity by the north-south wave motions and redistribution of wave vorticity by the zonal flow. An expression for the wavelength of a stationary disturbance satisfying these dynamics was derived, and slow changes in the longitude of centers of action were interpreted in terms of shifts in the mean zonal winds.

Quantitative modeling of the longitudinal variation of quasi-stationary patterns requires both an adequate atmospheric dynamic model and parameterization of the orographic and thermal forcing. *Charney and Eliassen* [1949] first attempted a quantitative model assuming the simplest possible dynamics of a constant zonal flow and barotropic disturbances depending only on longitude. Orography was parameterized in terms of a vertical motion produced by the zonal flow forced up and down over bottom boundary slopes. With the addition of Ekman friction at the bottom boundary they derived a mean January 500-mbar height profile (Figure 6) in remarkable agreement with the observed longitudinal variation.

The longitudinal atmospheric variations due to forcing by variations of sensible and latent heating, however, are generally of the same magnitude, as was first discussed quantitatively by *Smagorinsky* [1953] in terms of a two-layer baroclinic model. There has since been published an extensive body of literature attempting to refine further the dynamic model and the physical parameterizations [e.g., *Saltzman,* 1965; *Sankar-Rao,* 1970], but the relative contributions by orographic and thermal forcing to the observed longitudinal asymmetries are still somewhat obscure, probably in part owing to inadequate parameterization of the role of orography, as discussed for example by *Saltzman and Irsch* [1972]. More recently, *Manabe and Terpstra* [1974] have discussed various theoretical studies on this subject in the light of the results from their numerical experiments (see also our section D3*d*). The importance of developing further the description of the coupling between atmospheric heat sources and surface conditions has been recognized for some time [e.g., *Döös,* 1962].

The idea that perturbations in the atmospheric pressure pattern at one location can propagate great distances in latitude and longitude through the mechanism of planetary (Rossby) waves has provided a conceptual basis for statistical procedures now used for 'long-range forecasting' in the United States [e.g., *Namias,* 1968; *O'Conner,* 1969]. *Namias* [1971, 1972] has presented extensive evidence that slow fluctuations in the pressure pattern and hence seasonal anomalies

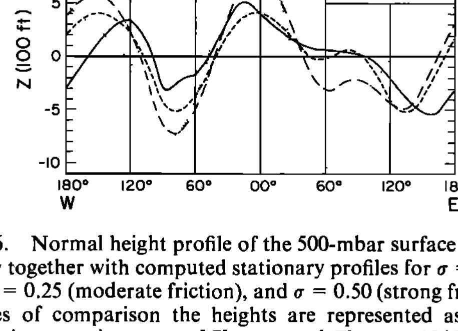

Fig. 6. Normal height profile of the 500-mbar surface at 45°N for January together with computed stationary profiles for $\sigma = 0$ (no friction), $\sigma = 0.25$ (moderate friction), and $\sigma = 0.50$ (strong friction). For purposes of comparison the heights are represented as deviations from their respective means [*Charney and Eliassen,* 1949].

in temperature and precipitation over the United States are significantly correlated with anomalies in the Pacific sea surface temperatures. Thus a forecasting procedure for anomalies in the atmospheric circulation is suggested; i.e., slow variations of sea surface temperature are predicted, and past observations are used to correlate these sea surface temperature anomalies with atmospheric circulation patterns.

e. Zonally symmetric models of the earth-atmosphere system. It is ultimately desirable to couple atmospheric models that include detailed description of zonal dynamics as a function of latitude and altitude to detailed models of the zonal surface energy balance. Necessary parameterizations include the relationship of the vertical structure of cloudiness, humidity, and latent heat release to the mean motion and temperature fields; the dependence of the model cryosphere on the model hydrological cycle; adequate description of the relationship between eddy heat and momentum fluxes and the zonal mean state; and the dependence of ocean temperatures on the surface heat budget and the zonal mean atmospheric state. The most ambitious model along these lines to date is being developed by *MacCracken* [1973].

f. Three-dimensional circulation models of the atmosphere. There are conceptually two complementary approaches to modeling the three-dimensional mean atmospheric circulation. The first approach involves simulating the atmospheric general circulation over a long time period with a GCM. We refer to such a model that prescribes ocean temperatures and sometimes snow and ice cover as an atmospheric GCM. The model is a time-dependent set of 'primitive equations' that describe the detailed evolution of the dynamic and thermodynamical state of the atmosphere. The resolution of the GCM is typically a few degrees of latitude and longitude horizontally and a few kilometers or so vertically. Processes occurring on scales larger than these are explicitly treated, whereas phenomena occurring on smaller scales (e.g., cumulus convection, latent heating, turbulent transport, and cloud formation) must be parameterized. Hydrological cycle components, radiative transfer, boundary layer processes, dynamic processes, and land surface temperatures are all computed simultaneously, as is described in the review by *Smagorinsky* [1974]. Atmospheric GCM's are currently able to simulate very well many of the observed large-scale features of the atmosphere, but most are still not properly 'driven' by the balance between incoming solar energy and outgoing infrared radiation. More correctly, most respond primarily to the infinite thermal capacity (i.e., fixed temperature) sea surface temperature boundary condition.

Our discussion of atmospheric GCM's in this article is brief, in part because of the availability of the review by *Smagorinsky* [1974]. We do, however, refer often to GCM's. The points that we do discuss and the models referred to should not be taken as a balanced review of GCM's. We have invoked GCM research primarily to clarify certain issues, and our choice of particular references is based only on our familiarity with the literature, not on any subjective judgment on our part of which particular model is best suited to clarify that issue.

An alternative, more mechanistic approach to three-dimensional modeling of the mean state would be to superimpose results from zonal mean and zonal asymmetric models as suggested by *Saltzman* [1968*a*]. Insofar as the asymmetries depend on the zonal mean state and vice versa, this becomes an iterative calculation. Because of the difficulties in using observations to prescribe the forcing for such models much benefit can be gained by using the GCM's to evaluate the forcing terms. The mechanistic models in turn provide diagnostic methods of interpreting climatic means from the GCM. Of course, the viability of such an approach has yet to be demonstrated conclusively in practice.

g. Three-dimensional circulation models of the earth-atmosphere system. It is essential for any attempts to predict long-period climate change that sea surface temperatures be calculated from model-generated heat balance requirements rather than prescribed ones and likewise that clouds be properly calculated rather than based on climatological distribution. Otherwise, a self-consistent global mean balance between incoming solar radiation and outgoing thermal radiation cannot be achieved. Sea surface temperatures would be derived most realistically in terms of a 'general circulation' ocean model driven by net thermal and momentum inputs at the interface. Much simpler ocean temperature prediction models, for example, a spatially dependent departure from global averages based on observed departures, may suffice for modeling global mean climatic variations over shorter periods of time, but their validity can be established only by reference to more realistic approaches. Cloudiness and snow and ice cover must also be parameterized in terms of model variables in order for long-term climate change to be investigated. A prescription of present ocean temperatures builds into an atmospheric GCM much of our present climate, as does a prescription of the observed distribution of albedo.

Manabe and Bryan [1969], at the NOAA Geophysical Fluid Dynamics Laboratory (GFDL) at Princeton University, have made first attempts at jointly modeling the atmosphere, ocean, and cryosphere. The GFDL atmospheric GCM [e.g., *Manabe et al.*, 1965] was coupled to an ocean model (a modified form of that described by *Bryan and Cox* [1968]) that computed three-dimensional fields of temperature, salinity, and horizontal motions explicitly and calculated density from a realistic equation of state. They used five vertical levels to the 'bottom' of the ocean and horizontal resolution comparable to that of the atmospheric GCM (except for greater resolution near western boundaries). A simplified land-sea distribution consisting of one continent and one ocean, a 120° longitudinal slice with cyclic continuity, and an annual mean heating distribution were used in this experiment. More recently, *Wetherald and Manabe* [1972] have used the same geometry to study the effect of seasonal changes over an annual cycle.

Since fluid motion in the atmosphere and fluid motion in the oceans occur on very different time scales, it was not computationally feasible in calculating the long-term adjustment of the ocean to operate the atmospheric and oceanic submodels on the same time scale. As described by *Manabe and Bryan* [1969], 'the atmosphere on the 0th, 0.5th, and 1st atmospheric years interacts with the ocean on 0th, 50th and 100th oceanic year of time integration.' The average temperature of the upper 50 m of the ocean is assumed to be associated with the constant temperature 'mixed layer' of the oceans, and the temperature of this layer is taken as the surface temperature boundary condition for the atmosphere. *Wetherald and Manabe* [1972] added more levels in the upper layer, and after using long ocean time steps as described above to reach equilibrium with annual mean heating, they ran ocean and atmosphere over the same time interval of an annual cycle.

The atmosphere provides the upper boundary condition, i.e., fluxes of heat, momentum, and moisture, to the ocean. In the Manabe-Bryan calculation, a 'running time-mean operator is applied to vertical fluxes (at the sea surface) to avoid an overresponse of the ocean model to features caused by individual synoptic disturbances in the atmospheric model' that otherwise (because of the disparate time scales between submodels) would be applied to the oceans without feedback for 100 times longer than would occur in reality.

Their results indicated important effects that depend on the interaction between jointly calculated atmosphere and ocean: for example, a drastic reduction of rainfall over the tropical ocean resulting from equatorial upwelling that altered the ratio of land-sea precipitation. Furthermore, even after computations for 100 years of simulated time the oceanic temperatures in the coupled model were still changing, suggesting that deep oceanic circulations could cause climatic fluctuations on this time scale.

Several interesting questions are raised by the Manabe and Bryan experiment. For example, what is the physical significance of the statistics generated by atmospheric and oceanic submodels coupled together with a 2-order-of-magnitude difference in the time intervals of the two models? These authors, pressed by computational necessities, have assumed in essence that the model's atmospheric state would be driven rapidly to an equilibrium forced by the slowly varying external oceanic boundary conditions and that this equilibrium state would subsequently provide mean flux conditions to the upper layers of the ocean that in turn would change its dynamic conditions, leading to a new set of atmospheric equilibrium fields. This approach implies that the atmosphere itself is transitive or ergodic (as discussed in sections A2–5) and will have a definite equilibrium state, given the external oceanic boundary forcing. (It is interesting to speculate on how the GFDL atmosphere-ocean model would behave if its climatologically specified cloudiness were replaced with cloudiness derived from a parameterization coupled to the hydrological cycle. In particular, would cloudiness coupling make the model atmosphere intransitive? See also the discussion in section B2*g* on temperature lapse rate feedback.)

3. *Modeling Conditions at the Lower Boundary*

a. Role of the lower boundary in climate modeling. Those aspects of climate that have the greatest societal importance largely involve conditions at the interfaces between the atmosphere and the land, sea, or ice. Surface climatic variables, especially temperature, have been monitored throughout much of historical time and are the most accessible parameters for studying past climates by means of geological, biological, and sociological records. The surface heat budget models of Budyko and Sellers have indicated a large sensitivity of surface temperature to changes of a fraction of a percent in net incoming radiation. It is thus important to recognize and to include in climate models all surface processes either that could vary enough on a long time scale to change the global mean surface heat budget by more than, say, 0.1%, or alternatively that could provide sufficient feedback on surface temperatures to alter the surface temperature changes (relative to those that might be calculated without that feedback) by more than, say, 10%.

It is sometimes useful to regard net radiative imbalance at the surface as the driving force for nonradiative processes, especially for situations where the imbalance is specified in terms of given surface temperatures. The net imbalance R consists of direct and diffuse solar radiation terms minus the net infrared loss ΔI:

$$R = Q(1 - \alpha) + Q_d(1 - \alpha_d) - \Delta I \qquad (2)$$

where Q is the unscattered part of the beam of solar energy reaching the surface, Q_d is the diffuse solar radiation flux reaching the surface, and α and α_d are the albedos of the surface for direct and diffuse radiation, respectively.

Obviously, surface albedo is a very important climatic parameter, and climate will be sensitive to changes of albedo greater than a few tenths of a percent on the global average. The next section treats in greater detail the different processes that affect global albedo and its coupling to other climatic variables.

The net radiative imbalance at a surface and the sum of all other energy transfer processes must balance the net rate of energy storage in a vertical column from the surface down to a level where there is no significant heat transport, denoted as G:

$$G = R - E_s - E_l - \Delta H + S \qquad (3)$$

where E_s is the sensible heat transported vertically away from the surface, E_l is the latent heat energy transported from the surface upward into the atmosphere, ΔH is the net heat transported horizontally out of the column, and S refers to all other surface energy processes, usually considered second order for long-term averages (or averaging to zero, since they are seasonal). Some examples of processes usually regarded as being second order globally are (1) snowmelt (which *Sellers* [1965] says is about 5–10% of the radiation balance in midlatitudes in the spring), (2) dissipation of mechanical energy by winds, waves, and tides, (3) heat transfer by precipitation, (4) photosynthesis, (5) oxidation of biological material (which can be very large locally), and (6) fires, volcanoes, and heat released by man's activities.

Sensible and latent heat as well as momentum are transferred from the surface to the atmosphere primarily by small-scale turbulent motions. The parameterization of this upward flux (and especially the dependence of this upward flux on other climatic variables) is an important aspect of climate modeling and is further considered in section C3*c*. The downward flux of heat and momentum into the ocean is almost equally important insofar as it determines to a large extent ocean surface temperatures. It can be treated by similar techniques. The processes determining ocean surface temperatures are further reviewed in section C3*d*.

It is found over land that the relative magnitudes of upward fluxes of sensible and latent heat are quite sensitive to soil moisture, the ratio E_s/E_l ranging from 0.1 over a water surface to about 0.5 over vegetated areas to 10 or greater over dry or desert areas [*Budyko,* 1958; *Sellers,* 1965; *Mitchell,* 1971]. This ratio $B = E_s/E_l$ is referred to as the Bowen ratio. For the earth as a whole, *Sellers* [1965, p. 104] asserts that 'evaporation accounts for 82% of the net radiation and turbulent heat exchange for 18%, and thus the main method by which the radiative heat surplus of the earth's surface is dissipated and transferred vertically to the atmosphere is by evaporation of water.' Table 1 [after *Sellers,* 1965] illustrates mean values of these terms, which on the average sum to the radiative imbalance R.

In polar latitudes and middle latitudes during winter the single most important climatological variable for determining surface albedo is the extent and nature of ice and snow formation. Since this formation in turn is very dependent on other internal climatic variables, these processes must be satisfac-

TABLE 1. Annual Energy Balance*

Area	R	E_l	E_s	E_s/E_l
Land	49	25	24	0.96
Oceans	82	74	8	0.11

* In kilolangleys per year.

torily parameterized in climate models. This topic is treated in section C3*e*. The question of parameterizing soil moisture for climate models is addressed in section C3*f*.

b. Albedo. The global mean solar radiation reflected back to space by clouds, aerosols, and surface conditions is given by a planetary albedo of approximately 0.30 [*Vonder Haar and Suomi,* 1971]. Our knowledge of the space and time distribution of the microscopic optical parameters of clouds and aerosols is still too primitive to motivate detailed radiative calculations based on these distributions in large-scale models. However, the effect of feedbacks between cloudiness distribution, surface albedo, and other internal variables (as discussed in section B2) can be evaluated to first order by using simple cloud models, cloud radiative effects being parameterized in terms of a mean reflectivity, absorptivity, and transmissivity most simply assumed constant. More correctly, these parameters should depend on cloud thickness, liquid water content, and water vapor concentration. A parameterization including some of these dependencies has been developed by *Lacis and Hansen* [1974] for the Goddard Institute for Space Studies (GISS) GCM [*Somerville et al.,* 1974].

From the standpoint of surface climate we are primarily interested in the solar radiation that reaches the surface, that is, the unattenuated (direct) solar radiation Q and the net forward scattered (diffuse) radiation Q_d in (2). The solar radiation absorbed in the atmosphere will affect the surface temperature only insofar as it (1) increases atmospheric temperature and hence the downward infrared radiation reaching the surface and (2) affects the convective stability of the atmosphere, which partially controls the vertical transports of sensible and latent heat from the surface. The increase in the downward flux of infrared radiation reaching the surface as a result of increased atmospheric absorption of solar radiation is generally considerably less than the increment of absorbed solar energy required to produce it and depends on the altitude of the incremental absorption. The net forward scattered radiation reaching the earth differs in general from unattenuated radiation in its mean angle of incidence. This difference in incident angle may be of climatic significance, as we shall discuss.

Variation of albedo with surface conditions is illustrated in Table 2 [after *Sellers,* 1965]. (The range of values given by *Sellers* [1965] for sea ice is lower than that generally observed [*Schwerdtfeger,* 1970; *Langleben,* 1971] and must apply to summer ice with extensive puddling.) To a lowest approximation we can distinguish between the albedo of the oceans, that of dry land surfaces, and that of snow-ice surfaces. However, it is apparent from the variations of albedo suggested by Table 2 that at least for those surfaces that are global scale in area, more refined distinctions are necessary, especially when variation of albedo with time is possible.

In particular, we infer from Table 2 that surface characteristics changed through human activity can typically change surface albedo by 10%. If such changes were to affect 10% of the continental area (3% of the global area), they could be of more than regional climatological significance. Also evident from Table 2 is the strong dependence of surface albedo in middle latitudes on the extent of snow cover. *Sellers* [1965] states that at least 60% more solar energy will be absorbed by the land surface between 40° and 45°N in a mild winter than in a cold, wet winter. This provides a positive feedback for maintaining anomalous winter conditions [*Namias,* 1962].

It is not possible to treat the contribution of the surface albedo to the planetary albedo independently of aerosols, cloud droplets, and molecules, whose albedos increase with increasing solar zenith angle. At sufficiently large solar zenith angles the solar radiation must penetrate an essentially infinitely scattering atmosphere that obscures the surface, whereas at low solar zenith it penetrates more easily to the surface. Consequently, at large zenith the fractional change of planetary albedo resulting from a change of surface albedo becomes significantly less than the fractional change of surface albedo [*Coakley and Schneider,* 1974]. In high latitudes this zenith angle effect may significantly reduce the temperature dependence of the planetary albedo from values inferred from near-surface measurements and thus may alter the energy balance calculations discussed later in section D2.

It may be important in large-scale models to allow for subgrid scale exposed surfaces (not covered by snow), which occur with light mean snow cover. *Holloway and Manabe* [1971] use a modified surface albedo A_b, given by

$$\begin{aligned} A_b &= A_g + S^{1/2}(A_s - A_g) \qquad S < 1 \text{ cm} \\ A_b &= A_s \qquad\qquad\qquad\qquad\;\; S \geq 1 \text{ cm} \end{aligned} \tag{4}$$

where A_g is the albedo of bare soil, A_s is the albedo of snow (assumed to be 0.60 for latitudes equatorward of 60° and to be 0.75 for latitudes poleward of 60°), and S is the liquid water content of the snow. They assume in using (4) that the ratio of liquid water to snow depth is about 1:10. Thus when the snow depth exceeds 10 cm, the surface albedo is independent of depth. They do not allow for a dependence on different types of vegetation, such as trees or forests, where the albedo dependence on snow depth would be quite different from that over a prairie or treeless region.

Table 2 indicates that the dependence of the sea surface albedo on solar zenith angle is such that the albedo of the sea surface increases with increased solar zenith angle. Clouds and other scatterers increase the fraction of diffuse radiation Q_d, which has a lower effective solar elevation angle than a high sun and a higher effective elevation than a low sun. Hence clouds increase the surface reflectivity when the sun's elevation is high (in the tropics) and decrease the albedo when

TABLE 2. Mean Albedos of Various Surfaces to Solar Radiation

Surface	Albedo, %
Water, plane surface	2.4
Water, at equator	6
Water, diffuse solar radiation	17
Water, 60° latitude in winter	21
Snow, fresh fallen	75–95
Snow, several days old	40–70
Ice, sea	30–40*
Soil, dark	5–15
Soil, dry light sand	25–45
Forest, coniferous	5–15
Crops	15–25

* Lower than generally accepted, as discussed in the text.

the elevation is low (at the poles). That is, $\alpha_d > \alpha$ for high solar elevation, and $\alpha_d < \alpha$ for low solar elevation (Table 2).

A related effect is the production of waves by surface winds, causing a sloping of the sea surface, and of foam, both having reflective characteristics different from those of a calm sea and both being dependent on solar zenith angle. *Kraus* [1972, chapter 3] discusses these effects in more detail. Through this dependence of surface albedo on surface winds, there is a feedback between the winds and the thermal balance, the outcome of which is not obvious.

One consequence of the zenith angle dependence of sea surface reflectivity shown in Table 2 is that an increase in cloud amount or man-made aerosols could result in an increase of surface albedo in tropical regions and a decrease in surface albedo in polar regions. Whether this could alter the equator-to-pole temperature gradient sufficiently to affect the generation of mid-latitude baroclinic disturbances and thus have a noticeable climatic impact is a question that deserves further study.

c. Parameterization of boundary layer transports. The treatment of atmospheric boundary layers has evolved from the classical theories developed by Prandtl, Taylor, Von Karman, and others for flow over a flat surface. Above the molecular sublayer there is found a surface boundary layer 10–100 m thick across which the vertical fluxes do not change significantly (as has recently been reviewed by *Kraus* [1972, p. 136]). Kraus estimates this layer in the ocean to be only 0.03 as thick as that in the atmosphere. The surface layer is considered to be stirred by eddies generated from the flow over the rough surface and depends on the classical 'roughness' scale z_0 and the Monin-Obukhov length L_s, a measure of the distance above the surface at which stratification effects become important.

Sources of convection sometimes produce a well-mixed layer to a height well above the surface layer. With neutral or stable stratification, complete mixing will be suppressed, but it is generally assumed that small-scale eddy motions still extend up to some level u_*/f, where f is the Coriolis parameter and u_* is the 'friction velocity' (ρu_*^2 gives surface stress). The region within which divergence of vertical momentum transport by small-scale eddies is comparable in magnitude to the Coriolis force on the mean wind is generally referred to as the Ekman layer and is characterized by large departures of the mean winds from geostrophic balance. The combined friction layer of constant shear stress and the Ekman layer make up the planetary boundary layer.

From the standpoint of climate modeling we need to parameterize in convenient form the vertical fluxes through these layers averaged over time and space. The flux within the constant flux layer is usually related to a nondimensional 'drag coefficient' C_D times the quantity being transported times the velocity of the atmosphere, all being evaluated at some reference level (typically 10 m) within the constant stress layer; for example,

$$\begin{aligned} \tau &= \rho \mathbf{v} \cdot C_D \, |\mathbf{v}| \\ E_s &= \rho c_p (T_0 - T) \cdot C_D |\mathbf{v}| \\ E_l &= \rho (q_0 - q) L \cdot C_D |\mathbf{v}| \end{aligned} \tag{5}$$

In (5), τ is the constant layer shear stress, ρ the fluid density, $\mathbf{v}$ the velocity at a reference height of about 10 m, L the latent heat of evaporation, c_p the specific heat, T the reference level temperature, T_0 the surface temperature, q the reference level specific humidity, q_0 the surface specific humidity (which over a wet surface will be the saturation humidity at temperature T_0), and C_D the drag coefficient appropriate to the particular atmospheric quantities. Variants of these basic formulas are used in most dynamic models of the atmosphere. The drag coefficient C_D at 10 m has been determined experimentally to be of the order of 10^{-3}, although there may be some dependence on $\mathbf{v}$ and on the nature of the surface (see *Hicks* [1972] for discussion and further references). At present, it appears difficult to improve upon these formulas over the open ocean, and expected mean errors are less than 30% [*Kraus*, 1972, p. 164]. For levels well below the Monin-Obukhov length L_s the surface winds have the classical log profile

$$U(z) = k^{-1} u_* \log (z/z_0) \tag{6}$$

where k is the Von Karman constant. Consequently, the drag coefficient can be extrapolated from 10 m to some other level z within the region of validity of (6) by using

$$C_D(z) = C_D(10 \text{ m}) \, [\log (z)/\log (10 \text{ m})]^{-2} \tag{7}$$

For models that essentially resolve the planetary boundary layer down to the constant stress layer it becomes possible to use (7) directly (or the appropriate generalization) to relate surface fluxes to wind at the lowest model level. This is attempted in the GFDL GCM models [e.g., *Holloway and Manabe*, 1971]. Otherwise, the structure of the planetary boundary layer must be regarded as determining surface fluxes in terms of variables calculated at the lowest model level. The NCAR GCM [e.g., *Kasahara and Washington*, 1971] parameterizes the transfer of heat, water vapor, and momentum in the planetary boundary layer in terms of eddy mixing, the mixing coefficient depending on the Richardson number according to a semiempirically calculated expression. The variation of temperature, water vapor, and winds from the lowest model level to the surface is just that required to transfer the fluxes from the surface to the lowest layer by eddy mixing. For assumed finite differencing, this balance provides values of parameters at the surface and hence surface fluxes. *Deardorff* [1972] has suggested a more elaborate parameterization based on detailed boundary layer calculations, where surface parameter values and hence fluxes are found graphically in terms of lowest model level variables, a bulk Richardson number for the planetary boundary layer, and the height of the planetary boundary layer.

One difficulty that should be noted in the application of (5) to derive fluxes for a long-period climate model is the inappropriateness of using time-mean parameters. For example, *Gavrilin and Monin* [1970] have shown for the GFDL model that much of the mean heat and humidity exchange may occur (at least at some locations) within transient storms. It is likely that velocity variances as well as mean velocities are required to evaluate adequately the stress from (5) in a time-averaged model.

Another problem in using the usual boundary layer formulations is that they do not recognize possible sub-grid scale but nonlocal transfers of momentum by wave motions (see, for example, the brief review by *Lilly* [1972] of present work on this problem).

d. Factors controlling ocean surface temperature. The net addition of radiation to the ocean, averaged over the globe for a long enough time, closely balances the fluxes of latent and sensible heat from the ocean surface into the atmosphere (Table 1). The oceans are much more opaque than the at-

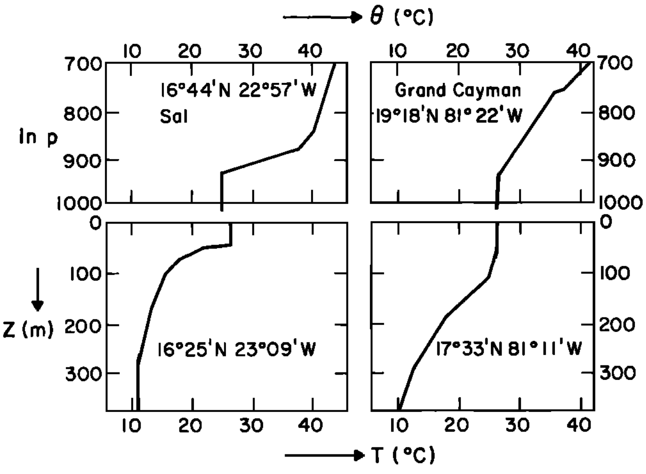

Fig. 7. Profiles of atmospheric potential temperature $\theta(\ln p)$ and oceanic temperature $T(z)$ in adjacent locations in the eastern tropical Atlantic and in the Caribbean [*Kraus and Turner,* 1967].

mosphere to visible and infrared radiation, so that radiative sources and sinks occur essentially at the ocean surface. The top layers of the ocean are completely mixed by the turbulence associated with generation of surface waves. In winter this mixed region is deepened significantly by downward convection of dense water formed by cooling and ice formation at the surface. At the base of the mixed layer the temperature drops to its value at the top of the stably stratified underlying water (Figure 7). The mixed layer grows in depth by 'entraining' the colder water at its base. The seasonally varying interface between the bottom of the mixed region and the top of the underlying stratified region is known as the 'seasonal thermocline.'

Surface heating depends on temperature through the temperature dependence of its infrared emission and through sensible and latent heat fluxes. Net heat input is very quickly distributed in the vertical direction throughout the mixed layer depth. The temperature of the mixed layer as a whole adjusts to this net heating (or cooling) and entrainment of fluid from below by increasing (or decreasing) until thermal balance is achieved.

The observed seasonal structure at two midoceanic stations is illustrated by Figure 8 [after *Kraus and Turner,* 1967]. Net heating during the spring forms a warm layer at the surface. This heating increases the temperature and decreases the depth of the mixed layer going into the summer months. With net cooling in the fall, the depth of the mixed layer increases rapidly until it has dropped to a depth of the order of 100 m (or more) by midwinter. As the annual cycle begins again and another warm layer forms at the surface, the ocean retains the memory of its past heating down to the greatest depth that the seasonal thermocline reached the previous winter (i.e., the top of the permanent thermocline).

The variation in depth of the seasonal thermocline is primarily a function of the amplitude of the seasonal cycle of incoming solar energy. Thus one would expect a rather permanent vertical structure in tropical latitudes, where the seasonal variation in solar input is relatively small, and an increasing amount of seasonal variation in the structure of the upper layers of the ocean with latitude, since the amplitude of the seasonal cycle of solar input increases with latitude. This expectation is more or less confirmed by observations. However, as was pointed out by *van Loon* [1966], there are exceptions. For example, even though the seasonal variation in incoming solar radiation is greater at latitude 50°S than at 35°S, the observed 'annual surface temperature range is smaller at latitude 50°S than at latitude 35°S.' Van Loon attributes this to the influence of 'clouds and the mixing of heat to greater depths at latitude 50°S (presumably due to stronger winds).'

To derive a simple estimate of seasonal temperature fluctuation, consider the thermal response of a mixed layer of depth h:

$$\rho C_p h \; \partial T/\partial t = Q_0 \cos(\nu t)$$

Assume a typical middle latitude seasonal fluctuation in net thermal input of $Q_0 \sim 200$ cal cm^{-2} d^{-1}. Then with $\rho C_p = 1$ in these units and $\nu = 2\pi/(365 \text{ days})$ we find for the amplitude of a seasonal temperature fluctuation

$$\Delta T \approx \frac{200 \times 365}{h} \approx \frac{100°\mathrm{K}}{h(\mathrm{m})} \tag{8}$$

giving a 10° amplitude for a shallow summer mixed layer 10 m deep and a 1° amplitude for a typical winter mixed layer 100 m deep. These values roughly correspond to the temperature variations indicated in Figure 8 (i.e., monthly evolution of changes in T at a depth of about 20 feet (6 m)).

On time scales longer than an annual scale, changes in thermal input will change the temperature of the mixed layer by an amount inversely proportional to the thermal relaxation rate of a column, that is, the rate at which the column temperature is adjusted to achieve thermal balance as a result of the temperature dependence of the heating rates. Again, this relaxation rate, and hence the change in temperature, is inversely proportional to the mean depth of the mixed layer, as is true in (8) but with a different proportionality constant.

Wind stresses on the surface also influence the thermal balance of the mixed layer by driving mean Ekman drift currents, which advect thermal energy horizontally and force

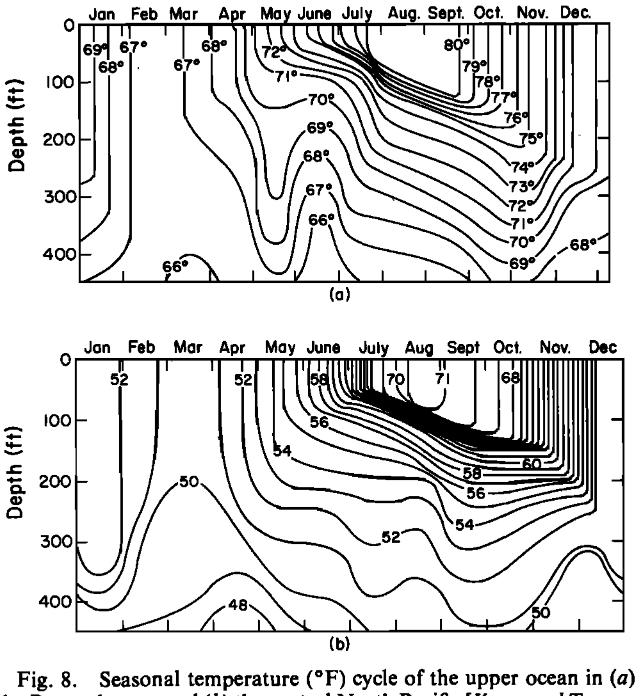

Fig. 8. Seasonal temperature (°F) cycle of the upper ocean in (*a*) the Bermuda area and (*b*) the central North Pacific [*Kraus and Turner,* 1967].

a mean vertical motion at the base of the mixed layer.

It is generally believed, however, that much of the global scale net horizontal transport of energy by the oceans occurs below the level to which the seasonal thermocline penetrates. The details of this horizontal transport are poorly known, and the vertical energy transfer into this region from above is even more poorly understood. Vertical transfer is often parameterized by eddy mixing and large-scale advection. For an eddy mixing coefficient K of 1 cm^2 s^{-1} and a depth of 200 m the time scale for changing the temperature of the underlying layers is $t_c \sim h^2/K \sim 1$ decade, so that for studying fluctuations on the time scale of a year it may be sufficient to take the temperature of the underlying layers to be given. The equator-to-pole energy transport on shorter time scales is then represented by some kind of vertical diffusion of energy between the mixed layers and the underlying layers, downward in low latitudes and upward in high latitudes, the horizontal advection being assumed to occur on longer time scales than are treated by the model. A more detailed description of this horizontal transport is then required for modeling climate change on longer time scales.

In summary, modeling ocean surface temperatures on time scales of a few years or less necessarily requires an adequate parameterization of the surface mixed layers; for longer time scales the deep circulation must be treated as well. We shall now briefly summarize some of the recent progress in these two areas.

The pioneering model of the surface mixed layer is that of *Kraus and Turner* [1967]. When complete mixing is assumed, the mixed layer is described simply by its temperature T_s and its depth h. Two equations are derived that govern these parameters. The first relationship is that the temperature changes because of added heat or entrainment of colder water from the underlying seasonal thermocline. The second, less straightforward relationship is an equation for conservation of kinetic and potential energy. The kinetic energy is that of turbulence within the mixed layer that is generated by wind acting on waves at the air-sea interface. The kinetic energy storage is neglected, so that its generation is balanced by dissipation or conversion into potential energy. Potential energy in turn changes with variations in the temperature or depth of the mixed layer; thus two equations for h and T_s are derived. *Denman and Miyake* [1973] have related the theory to diurnal and day-to-day variations by using 2 weeks of temperature data from ocean station 'Papa,' together with refinements of the theory made by *Denman* [1973].

The original Kraus-Turner model assumes that wind blowing over the ocean surface generates turbulence, but that model neglects any currents set up by the wind stress. *Pollard et al.* [1973] have shown, on the other hand, that if the generation of turbulent kinetic energy is neglected but the generation of kinetic energy of wind-driven inertial currents is allowed for, a very different description of the time evolution of the mixed layer is derived. More recently, *Niiler* [1974] has reconciled these two approaches by including generation of both turbulent and mean current kinetic energy and has shown that in the absence of heating, this combined model reproduces the results of the Kraus-Turner model for time scales that are large in comparison with an inertial day. For intermediate time scales, inertial motions play an important role in his model. Pollard et al. showed with their model that the inertial motions could drive the mixed layer to some constant depth in a time of the order of an inertial day. In Niiler's model if the mixed layer is shallower than this constant depth, the excitation of inertial motions by wind stress still drops the mixed layer to this level in a period of the order of an inertial day. This level is thus interpreted as a minimum mixed layer depth.

The Kraus-Turner model of oceanic vertical mixing simulates 'penetrative convection'; that is, the region of mixing entrains underlying stable fluid through the conversion of turbulent kinetic energy into potential energy. In this respect it differs from the more classical convective adjustment used for atmospheric temperature structure in climate models (as discussed in section D1*b*), where the condition assumed is not conservation of kinetic plus potential energy but rather absence of a discontinuity in temperature across the interface between the mixed and the stable region. *Gierasch* [1971] has discussed these two alternative approaches for atmospheric models and has carried out some calculations appropriate to the Martian troposphere to illustrate these differences in determining an atmospheric temperature profile.

The distinction between the two alternative approaches to mixing appears much more crucial for calculation of oceanic than of atmospheric temperature. The classical convective adjustment gives no oceanic mixed layer under conditions of net surface heating. Thus in this case there would be no wind mixed layer thermal inertia for surface heating, even though in reality the immediate thermal inertia of the ocean surface for heating depends entirely on the depth to which it has mechanically been stirred. With strong enough surface cooling the two approaches should give results that do not differ greatly in the response of the surface temperature to net heating.

Bryan [1969] and *Wetherald and Manabe* [1972] allowed only for instantaneous mixing due to negative buoyancy (i.e., classical convective adjustment). Otherwise, mixing was parameterized by turbulent diffusion with an eddy mixing coefficient of magnitude 1.0 cm^2 s^{-1}. Wetherald and Manabe were, however, able to model some features of the seasonal variation of sea surface temperature in middle latitudes. Their study suggests that these features have important climatic effects. The wintertime deep mixing by convective adjustment greatly increases the thermal inertia during that cooling season and causes a consequently smaller decrease of surface temperature. On the other hand, confinement of the heat added during the summertime to the upper layers allows a large increase in temperature. This effect is shown schematically in Figure 9. Hence depending on whether convective mixing of the surface layers is present or absent in an annual mean model, a model with seasons will give considerably warmer or cooler annual mean surface temperatures, respectively, at a given latitude than an annual

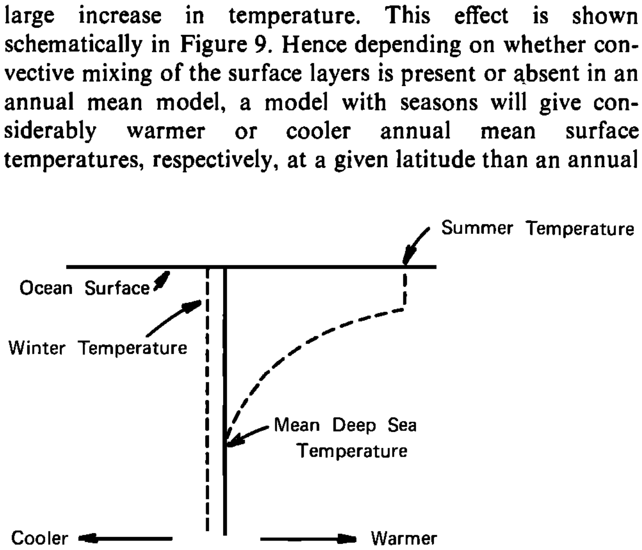

Fig. 9. Schematic diagram showing how the ocean is heated or cooled at high latitudes [*Wetherald and Manabe*, 1972].

mean model. Thus Wetherald and Manabe obtained in middle to high latitudes a significantly warmer annual mean ocean temperature than Bryan did. This effect (together with summertime removal of the snowpack and consequent albedo reduction) gave a significantly warmer mean climate in high latitudes. It is, however, not clear how much of this difference would be retained with a very long time integration of a dynamic ocean model. The answer to this question will suggest a minimum time step for integration of a climate model that includes a mixed layer ocean. In fact, a recent calculation by J. Miller (private communication, 1974) at GISS using an updated version of the mixed layer model of Denman showed that different mean surface temperatures could be obtained from model integration over a few-day period by using a diurnal solar cycle as opposed to a fixed (but equal in the time mean) value of solar input. If this result were true for equilibrium surface temperatures (i.e., after a few months or more of integration), it would imply that the minimum time step that should be used in a climate model is much less than 1 day! This possibility obviously needs further exploration.

In the study of Wetherald and Manabe, climate is presumably made warmer by an increased transfer of heat from the ocean surface, which must ultimately be balanced by some combination of increased poleward transport of thermal energy through oceanic circulation, and adjustment of interior ocean temperatures to remove the increased cooling from the ocean surface. Nevertheless, even if the annual cooling from the surface were to remain the same as it is in the annual model, there could be a significant climate change because of the changed partitioning of the net energy loss in the seasonal model into sensible, latent, and infrared cooling (all nonlinear processes), given a seasonally varying sea surface temperature rather than a mean value.

The strong dependence of the mixed layer depth on surface wind during the summer introduces another feedback between the ocean and the atmosphere whose effects have not yet been explored in the context of climate models. The implication is, however, that the summer ocean surface temperature will increase less than it might otherwise because of this wind stirring and that the more the stirring, the less the annual range as suggested by *van Loon* [1966], for example, as a partial explanation of the relatively small annual range at 50°S. Likewise, the effect of seasonal heating variation discussed by Wetherald and Manabe should be reduced with increasing wind strength.

So far, we have neglected the effects of salinity variation in our discussion of surface mixing. The variations of buoyancy due to salinity variations can also control mixing depth, especially in high latitudes. Salinity in the surface layers may change owing to a difference between evaporation and precipitation as well as by ocean current transport, formation or melting of sea ice, and river runoff. *Bryan* [1969] attempted to include the effects of all these processes in his model. Further discussion of the role of convection in the formation of ice is given in section C3*e*.

It is apparent that a much more thorough observational description of the structure of oceanic surface layers than is presently available will be a necessary prelude to accurate modeling of the interaction between these layers and the atmosphere. Even more discouraging is the possibility of immediately modeling in an entirely satisfactory manner the transport of heat by internal oceanic circulation. We have a general idea of the constraints that such a model must satisfy, but we are entirely ignorant of important details that should be included for such a model to describe adequately the feedbacks required for studying climate change in the atmosphere-ocean system.

For an oceanic circulation to be driven by the equator-to-pole heating differential, cold water must be transported equatorward, and warm water poleward. The difference in temperature between the cold equatorward-moving water and the warm poleward-moving water is generally assumed to result primarily from their difference in depth. That is, transport is by meridional overturning, there being upward motion at latitudes where there is surface warming and downward motion where there is surface cooling and the net poleward mass transport occurring above the net equatorward transport. Longitudinal variation of the circulation is essential, however, to satisfy the constraint of vorticity balance. At latitudes of mean upward motion this motion produces in the upper levels a horizontal mass divergence (because vertical motion has to go to zero at the surface). This divergence implies a shrinking of vortex tubes and hence a reduction of the total vorticity of a fluid column. For an inviscid steady state in the interior, only the planetary component of vorticity can be lost, and this implies an equatorward flow in the interior. (Planetary vorticity is the Coriolis parameter, and 'interior' here means away from boundaries.) Thus there must be continental boundary currents that carry enough thermal energy poleward to compensate not only for cooling at the surface but also for the equatorward transport required in the interior. Through similar arguments it is established that the equatorward flow at lower levels must likewise occur in a boundary current at those latitudes where mean upward motion occurs in the interior. The sinking of cold water from the surface to maintain these currents is believed to occur in a relatively small area in high latitudes during winter, where convective mixing down to great depths is possible (see, for example, the discussion by *Stommel* [1966]).

To model the above processes in detail, it is necessary to understand how heat is transferred vertically by small-scale processes and how eddy motions dissipate the vorticity of currents at continental (generally oceanic western) boundaries. These are perhaps the two fundamental questions in physical oceanography. Extensive research efforts have not yet yielded much more than a greater appreciation of the difficulty of these questions. The quantitative exploration of large-scale circulation has thus had to assume vertical diffusion and boundary layer dissipation as adjustable parameters prescribed in terms of eddy mixing coefficients. Such models have been reasonably successful in simulating analytically the permanent thermocline [e.g., *Veronis*, 1969; *Barcilon*, 1971; *Welander*, 1971]. The difficulties in initial value numerical integration of such models for an arbitrary basin are largely technical, i.e., the difficulty in achieving adequate resolution, especially in the western boundary layers, and the long thermal response time for the deeper layers.

Besides *Bryan*'s [1969] calculation, *Holland and Hirschman* [1972] have integrated a detailed model for circulation in the North Atlantic, density distribution being prescribed rather than calculated. This and other recent studies have indicated that proper inclusion of bottom topography may be essential to derivation of oceanic circulations. Besides the thermally driven circulation emphasized above, oceanographers also study the mean circulation driven by winds at the ocean surface. This circulation affects the ocean thermal fields through their geostrophic adjustment to balance the wind-driven currents, and it may be an important mechanism for redistribution of thermal energy in the surface layers on less than global scales. An understanding of the significance for climate of the coupling between the wind-driven and the thermally driven circulations awaits better general knowledge of large-scale ocean circulation. A recent review of numerical modeling of ocean circulations has been given by *Marchuk et al.* [1973].

e. Ice and snow. We discussed earlier the role of the cryosphere in modifying surface albedo and hence surface energy balance (sections B2*c* and C3*b*). It is useful to distinguish between three components of the cryosphere: (1) the massive continental ice caps of Antarctica and Greenland, which are generally believed to vary on a time scale of millennia, (2) the sea ice cover, which fluctuates to a great extent seasonally and whose distribution is apparently quite sensitive to climate change on a time scale of a decade or less, and (3) the seasonal snow cover, which during winter extends over continental areas into middle latitudes. (Here we do not consider ice in the atmosphere as part of our definition of the cryosphere.) The antarctic and Greenland ice caps cover 3 and 0.4% of the globe, respectively, and contain enough water to raise the mean sea level by about 59 and 6 m, respectively [*Flint*, 1971, p. 84].

In the southern hemisphere the distribution of sea ice relative to the global area varies seasonally between an astonishing 5% during winter to less than 1% during summer, whereas in the northern hemisphere the variation is only between 3% and 2% [*Fletcher*, 1969]. If this observed distribution of sea ice (covered by snow in winter) were replaced by water, the global albedo would be reduced by an amount of the order of 1%. Northern hemisphere continental snow cover extends over 7% of the globe during January [*Kukla and Kukla*, 1974]. (There is very little seasonal variation of snow cover in the southern hemisphere except on the sea ice.) Thus the mean net global albedo increase contributed by seasonally varying continental snow cover is comparable to that provided by the sea ice. *Vowinckel and Orvig* [1970] and *Schwerdtfeger* [1970] have reviewed the physical climatology of arctic and antarctic regions.

The formation of ice and snow depends on surface temperature, precipitation, and other factors that determine the surface energy budget and hydrological cycle. Thus a self-consistent climate model should determine the ice and snow cover as an internal degree of freedom of the model. Up to the present this has been done with very few models, for example, in a highly parameterized way in the energy balance models of *Budyko* [1969] and *Sellers* [1969, 1973] and in the GCM simulations of *Manabe and Bryan* [1969] and *Wetherald and Manabe* [1972].

The approach taken by Budyko and Sellers was to relate ice cover to temperature empirically so that the surface albedo was temperature dependent. Budyko assumed that at all latitudes where surface temperature dropped below the present mean temperature at 72°N the albedo of the earth-atmosphere system would increase from 0.50 to 0.62. Similarly, *Sellers* [1969] correlated albedos in the two hemispheres near the ice boundary with surface temperatures and inferred that at annual temperatures below 10°C the albedo would increase at a rate of 0.9% per 1°K drop in zonal mean annual temperature until a maximum value of 85% was reached. The predictions by the Sellers-Budyko models of variation in ice cover with net solar heating greatly stimulated interest in the coupling of ice albedo to global climatic variations. However, in view of the ad hoc manner in which ice was coupled to albedo, it is difficult to obtain quantitative verification of their conclusions even for very small departures from current conditions. Nevertheless, these models do show the first-order importance of albedo-temperature coupling on climatic stability (see sections C3*b* and D2 for further discussion).

A physically much more satisfactory approach is to derive the ice and snow cover from the hydrological cycle as attempted by Manabe and his co-workers. Their models distinguished snowfall from rainfall according to air temperature at a height of 350 m according to an empirical relationship. It seems unlikely that their parameterization of snowfall could be much further improved without a thorough overall improvement in their cloud physics parameterization. The accumulation of snow at the surface S is calculated in their model according to a budget equation $\partial S/\partial t = S_F - E - M_e$, where S_F is rate of snowfall; E is rate of evaporation; and M_e is rate of melting, which is determined from a heat balance condition at the surface. (The latest numerical study using these parameterizations is mentioned in section D3*d*.)

More detailed modeling of the snow cover requires consideration of snow temperature and radiative fluxes as functions of distance beneath the snow surface. Such multilayer models have been developed, for example, by *Maykut and Untersteiner* [1971] for the snow cover overlying the arctic pack ice and by *Schlatter* [1972] for selected locations in Antarctica.

New sea ice develops in high latitudes during the wintertime by cooling of the ocean surface to freezing temperatures followed by further surface heat loss. Once it is formed, ice accretes on the lower side of the sea ice during net energy loss and ablates largely from the top during net heating. The rate of accretion at the bottom depends on the rate at which heat can be conducted vertically upward through the ice to the top. Hence the rate of ice formation is very sensitive to the thickness of the existing ice. Indeed, its value in winter varies by 2 orders of magnitude as the ice thickness ranges from 0 to 3 m [*Maykut and Thorndike*, 1973]. *Maykut and Untersteiner* [1971] studied a model of the seasonal variation of a slab of sea ice, given the various energy fluxes at its top surface and the net loss of heat from the underlying ocean. They assumed a seasonal variation of surface albedo ranging from 0.64 for midsummer bare ice to 0.85 for midwinter snow cover. With their best estimates of appropriate boundary conditions they calculated (in agreement with observations) an average ice thickness of 3 m and an annual amount of ice ablation from the top and net accretion at the bottom of 40 cm. The sensitivity of their results to variations in boundary conditions was examined. Reduction of their summertime albedo by 20% led to complete melting of the ice pack by the third summer. (*Campbell and Martin* [1973] have discussed the issue that oil spills may induce such an albedo change.) An increase of upward heat flux in the ocean from the standard value of 2 to 6 (in kcal/cm^2 yr) also melted the ice. The ice thickness was insensitive to seasonal snowfall of less than 70 cm, but it greatly increased for larger snowfall.

The approach of Maykut and Untersteiner does not by itself give a completely self-consistent model of sea ice because it neglects horizontal heterogeneities of ice thickness and ice transport as well as feedback from the atmosphere above and the ocean below. Modeling of each of these omitted features is now discussed in turn.

The arctic ice dynamics joint experiment has been developed especially to improve the observational understanding of the dynamics of arctic sea ice. As was stated by *Maykut et al.* [1972],

> Large sheets of sea ice are in an almost continual motion that causes fracturing, ridging, and the formation of regions of open water. Neither the physical mechanisms controlling these processes nor their role in maintaining the overall size of the ice cover is adequately understood, despite a long history of scientific research in the Arctic.

One basic problem to be studied is the derivation of constitutive laws relating large-scale ice deformation to external stress fields; another is the mechanisms of ice deformation. *Maykut and Thorndike* [1973] suggest an approach to the thermodynamics of the ice pack with small-scale inhomogeneities by using a statistical distribution of ice thicknesses. A consequence of the large-scale movement of the pack ice is the export of sea ice from the Arctic Sea. For example, *Koerner* [1973] estimates that 15% (by mass) of the ice cover is exported each year.

Bryan attempted to model sea ice on top of his ocean model as parameterized by two variables, temperature and thickness. His ice model allowed for accretion or ablation in terms of surface energy input and conductive coupling of the ice surface to the sea temperature. The ice thickness was also assumed to be governed by transport by ocean currents and a 'viscosity' law. Bryan's ice pack did not reach a steady state but melted before the end of his integration. This is not surprising in view of the previously mentioned results of Maykut and Untersteiner. The sensitivity of ice thickness to boundary conditions is such that tuning parameters within existing uncertainties can make the difference between obtaining an ice thickness much greater than that observed and obtaining a complete melting of the ice.

Bryan's model for pack ice was also coupled to Manabe's atmospheric model. However, atmospheric GCM's that specify a zonal climatological cloudiness rather than calculate it locally may be rather unsatisfactory for determining coupling between ice and atmosphere. With the large zenith angles in polar regions the details of cloud distribution become very important for determining the net solar radiation reaching the surface and over water (including puddles over ice) in determining the surface albedo, which is zenith angle dependent (Table 2). The net infrared radiative cooling is also sensitive to details of the distribution of clouds. In view of the Maykut and Untersteiner results that ice radiation balance parameters should be correct to 20%, better surface energy balances calculated from the zenith angle dependence of multiple scattering by clouds and particles may be required. Since the presence of a low stratus cloud deck over the Arctic Sea in summer is essentially a permanent climatological feature, the stratus and its effects on the local energy balance should be properly simulated in an atmospheric submodel.

Vowinckel and Orvig [1970] have discussed the observed distribution of clouds in the Arctic. In particular, there appears to be a strong correlation between surface moisture and formation of low stratus clouds. Since low stratus is the dominant cloud form in the arctic summer and is very prevalent, it may be important to have a cloud prediction scheme for low arctic stratus clouds in GCM's. At present, most GCM's do not have adequate vertical resolution to describe low stratus formation.

An essential ingredient of an ocean model underlying the ice pack is a surface mixed layer (as discussed in section C3*d*). At near-freezing temperatures the vertical distribution of salinity is at least as important as the temperature distribution in stably stratifying the ocean and hence in limiting the depth of the mixed layer. If salinity and therefore density increase downward in arctic waters, it is necessary only to cool a top layer of water to freezing temperature before ice formation can begin; whereas with a uniform salinity distribution it might be necessary first to cool the water down to the bottom. Beneath the ice the processes that remove heat from the mixed layer and ultimately from the overlying ice presumably also depend on the mixed layer depth.

In the process of ice formation, only a small fraction of the salt is incorporated into the ice, so that brine is added to the sea, producing convection and driving the mixed layer down to greater depth. *Solomon* [1973] has examined how the mixed layer depth and salinity below freezing ice that are inferred from a classical convective adjustment calculation differ from those of the Kraus-Turner mixed layer model (see section C3*d*). It also appears necessary in view of Maykut and Untersteiner's results to model the export of heat from the arctic seas by large-scale currents.

The arctic pack ice is formed much further equatorward in the Pacific sector than in the Atlantic sector. *Weyl* [1968] interprets this zonal asymmetry as a consequence of the greater salinity of the North Atlantic near-surface waters; thus there must be deeper mixing of surface waters before freezing in the Atlantic than in the Pacific. He also argues that the more saline North Atlantic water spreads into the southern polar ocean and controls the extent of antarctic sea ice. Weyl then suggests that the differences between Pacific salinity and Atlantic salinity are controlled by net atmospheric transport of water vapor from the Atlantic to the Pacific (primarily across the Isthmus of Panama). He also hypothesizes that historically, when this net transport has been absent, there has been a marked increase in the equatorward extent of the sea ice, which if it were sufficiently persistent, could initiate an ice age. He argues that reduction of North Atlantic salinity could reduce the rate of bottom water formation in the Greenland-Norwegian Sea, thus reducing the strength of the Atlantic thermohaline circulation and hence providing positive feedback by reducing the northward advection of warm saline water from the North Atlantic into the region of bottom water formation.

Weyl's ice age hypothesis indicates how coupling between atmospheric dynamics, polar sea ice, and oceanic currents and salinity distribution may give rise to long-period internal oscillations in the climate system (such as are hypothesized in section A3). Much more modeling aimed at incorporating the coupling between these physical processes is required before quantitative validation of any such theories will be possible.

The continental ice masses of Antarctica and Greenland can vary substantially on time scales of the order of millennia and thus may be specified as given external conditions for calculations on shorter time scales. Models of continental ice sheet evolution become important, however, for consideration of major ice ages and interglacial periods. An ice sheet gains mass from net accumulation of snow that subsequently changes to ice in the climatologically colder (generally higher or more poleward) part of the sheet and loses mass in the climatologically warmer part of the ice. For a steady state the ice must then flow from the source to the sink region. Observational understanding of the governing flow laws is achieved through study of present-day glaciers. Imbalance between sources and sinks of ice leads to net growth or decay. Under certain conditions the flow laws can change drastically, leading to 'surges' of the ice mass. This can lead to a positive feedback effect, where the frictional drag between the ice and its lower boundary is decreased by a melting of the interfacial ice from heat generated by increased ice movement, which in turn leads to decreased drag and further slip.

f. Ground hydrology. The water vapor concentration over water and saturated land surfaces (giving the q_0 in (5)) depends only on surface temperature T_0. Hence for these surfaces and to the extent that $q - q_0$ in (5) depends only on q_0 and $(T - T_0)$ and, for example, can be written as

$$q - q_0 = (\partial q_0/\partial T)_{T_0}(T - T_0) \tag{9}$$

one can conclude that the ratio of vertical fluxes of sensible to latent heat depends only on surface temperature. The rate of evaporation that would take place from a saturated surface, known as the 'potential evaporation rate,' is generally used together with the moisture content of the soil for determining the rate of evaporation from unsaturated soil. *Priestley and Taylor* [1972] suggest an expression for this ratio to be used over saturated land surfaces, of the form

$$E_s/(E_s + E_l) = \alpha s/(s + \gamma) \tag{10}$$

where $\gamma = C_p/L$; $s = (\partial q_0/\partial T)_{T_0}$; and α is an empirical constant, equal to 1.26. Figure 10 [after *Priestley and Taylor,* 1972] shows the Bowen ratio E_s/E_l given by this relationship and some supporting data obtained over water surfaces. They caution that this expression will not always be applicable over water but suggest that it provides a useful basis for determining the rate of evaporation over a saturated land surface. For numerical modeling, potential evaporation is generally determined directly from (5), q_0 being given by the saturated humidity. (*Sellers* [1965] has reviewed other approaches that have been suggested for determining potential evaporation.)

It is generally assumed that a given soil type will hold a certain amount of water that is accessible to evaporation (or transpiration). When the soil moisture is greater than some critical value W_k, it is found that evapotranspiration proceeds at the potential rate E_0. A parameterization for evaporation

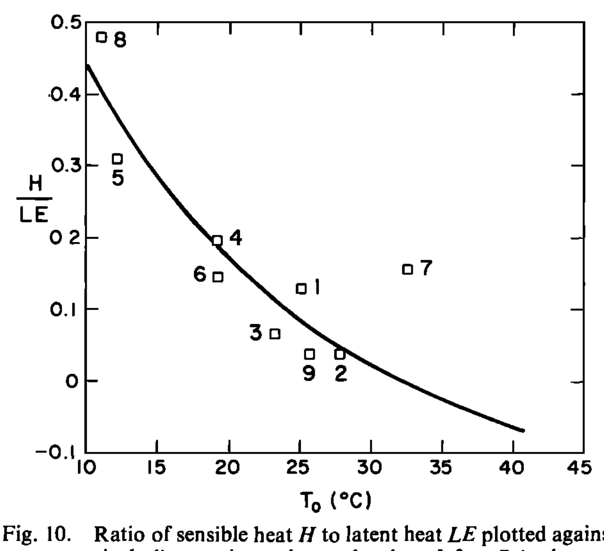

Fig. 10. Ratio of sensible heat H to latent heat LE plotted against temperature including various observed values [after *Priestley and Taylor,* 1972].

often assumed then is that

$$\begin{aligned} E_l &= E_0 & W &\geq W_k \\ E_l &= E_0 \cdot (W/W_k) & W &< W_k \end{aligned} \tag{11}$$

where W is the soil moisture content. Priestley and Taylor indicate that evaporation proceeds at the potential rate until all but approximately 5 cm of the available soil moisture content has been evaporated. That is, $W_k \simeq 5$ cm. The capacity of soil for moisture, W_{FC}, varies widely depending upon the kind of soil. Priestley and Taylor discuss different data that give W_{FC} between 5 and 20 cm.

One early attempt at introducing ground hydrology into a GCM was made by *Manabe* [1969]. He assumed that the soil everywhere had a field capacity of $W_{FC} = 15$ cm and $W_k = 0.75 \times W_{FC} \simeq 11$ cm. Manabe treats separately regions with or without snow cover. In the absence of snow cover,

$$\begin{aligned} \frac{\partial W}{\partial t} &= R - E & & W < W_{FC} \\ \frac{\partial W}{\partial t} &= 0 \qquad r = R - E & & W > W_{FC} \end{aligned} \tag{12}$$

where R is rate of rainfall, E rate of evaporation, and r runoff. He assumes that the snow holds zero moisture. Beneath the snow there is no evaporation, but snowmelt acts as an additional moisture source.

Manabe found in his model that allowing land to dry out affected the distribution of rainfall. In particular, in the subtropics he found that contours of small rainfall amount are more extended in latitude over continents than over oceans. He interpreted this effect as being due to positive feedback between rainfall and soil moisture and referred to it as 'a desert-forming mechanism.'

D. FURTHER DISCUSSION OF MATHEMATICAL ASPECTS OF CLIMATE MODELS

In previous sections we have outlined a hierarchy of climate models, have given a brief description of their features, and have cited many (but by no means all) references to existing work. Only in the section (C3) on modeling conditions at the earth's surface did we touch in any real detail on particular model parameterizations. In this section a more detailed but highly selective description of certain aspects of particular models is presented. General circulation models are not stressed because a thorough treatment of the formalism of these models is necessarily too lengthy for our present purposes and has recently been reviewed by *Smagorinsky* [1974]. Our purpose here is to elaborate on some of the basic assumptions of certain models that are important to climate theory.

1. *Horizontally Averaged Models*

Temperature is without doubt the one variable most generally regarded as being synonymous with climate. Earlier (equation (1)) we discussed the radiative equilibrium temperature of the earth as determined by a balance between the unreflected part of the incoming solar energy, $Q_s(1 - \alpha_p)$, and the net planetary infrared radiation, σT_p^4, escaping from the planet to space. Variations in the planetary radiative equilibrium temperature T_p are at best a crude measure of variations in the surface temperature T_s. In fact, as we shall see in section E1*a*, there may be circumstances when these parameters can vary in the opposite sense. Obviously, what is needed is a calculation of the vertical temperature profile, which, together with the vertical distribution and the optical properties of the optically active atmospheric constituents (such as CO_2, H_2O, O_3, clouds, and dust), determines the radiative fluxes (and therefore T_p) as well as the temperature at and adjacent to the surface.

Here we will examine models that compute the vertical distribution of radiative fluxes and temperature profile $T(z)$ for a column through the atmosphere. Horizontal energy exchange with neighboring columns is neglected. Such an approach is often most useful in application to a horizontal global average and can be used to identify the first-order effects of variations in radiative constituents of the atmosphere on the radiative balance and surface temperature.

The rate of change of atmospheric temperature at height z multiplied by the product of local density and specific heat, C_p, is given by the sum of the divergence of net (upward minus downward) infrared and solar radiative fluxes, $Q_{\text{net}} = F_{IR}\uparrow - F_{IR}\downarrow + Q_S\uparrow - Q_S\downarrow$, the vertical transport of sensible heat, and the release of latent heat of vaporization (the technique for modeling the latter two processes will be called 'convective adjustment'). The rate of temperature change is thus

$$\rho C_p \frac{\partial T(z)}{\partial t} = -\frac{\partial}{\partial z} Q_{\text{net}} + \text{convective adjustment} \qquad (13a)$$

a. Radiative flux calculation. To calculate the radiative heating rate term

$$\rho C_p \frac{\partial T_r}{\partial t} = -\frac{\partial}{\partial z} Q_{\text{net}} \qquad (13b)$$

requires knowledge of the vertical fluxes of solar and infrared radiation in terms of the temperature distribution and vertical distribution of optically active constituents.

Before proceeding further with the details of selected radiative transfer models we first briefly summarize (following *Schneider and Kellogg* [1973]) the processes of absorption of infrared radiation by gases such as carbon dioxide and water vapor. The transparency of these gases to solar radiation relative to their opaqueness to infrared radiation leads to the well-known greenhouse effect described qualitatively below.

The warm surface layer emits infrared radiation, most of which is intercepted by optically active atmospheric gases, clouds, and particles. These constituents reemit radiation both up to space and back down to the surface, the downward radiation reducing the net loss of heat from the surface. Since the atmospheric emitters are colder than the surface, they emit proportionally less radiant energy. Consequently, the total outgoing infrared radiation from the earth-atmosphere system is less than the radiant energy emitted by the surface alone, and the effective radiation temperature of the earth is influenced more by the temperature of the colder atmospheric gases and cloud tops (which emit radiation roughly like a blackbody with the temperature of the atmosphere at the cloud tops) than by the warmer surface below. Should the amount of an infrared-absorbing gas in the troposphere be increased, it would then intercept a larger fraction of the infrared energy coming upward from the warmer layers near the surface. Thus we have the greenhouse effect, by which the outgoing infrared flux to space is reduced and the downward infrared flux is increased by infrared-absorbing gases in the lower atmosphere. For the balance between infrared emission to space and solar flux absorption by the earth-atmosphere system to be maintained, on the assumption that the planetary albedo remains unchanged (but with an increase in the concentration of infrared absorbers in the atmosphere), the surface temperature must rise.

Next, we will consider briefly the equations of radiative transfer that describe the absorption of infrared radiation. The fundamental equation that describes infrared radiative transfer, obtained by using Kirchhoff's law and valid in the lower atmosphere to a sufficient degree of approximation [*Goody*, 1964], is

$$dI_\nu/ds = \rho_t K_\nu(I_\nu - B_\nu) \qquad (14)$$

where I_ν is the intensity of monochromatic radiation of wave number ν $(=1/\lambda)$ traveling along the path ds, B_ν is the intensity of blackbody or Planckian emission, ρ_t is the absorber density, and K_ν is the absorption coefficient per unit density. In the presence of two or more absorbers the absorption coefficient $\rho_t K_\nu$ should be replaced by the sum of the coefficients for each respective absorber. For a horizontally homogeneous atmosphere, (14) can be integrated over wave number and from the bottom of the atmosphere to the top and over all zenith angles in the form of an upward $F_{IR}\uparrow$ and downward $F_{IR}\downarrow$ radiation flux passing through the level $z = \bar{z}$.

The net infrared radiative flux in the atmosphere at level $z = \bar{z}$, F_{net}, is the difference between the upward and the downward fluxes, $F_{IR}\uparrow - F_{IR}\downarrow$, the vertical divergence of which is the infrared contribution to the rate used in (13). The upward flux at $z = \bar{z}$ for radiation at some frequency ν (in cm^{-1}) consists of two terms,

$$F_{\nu\uparrow}(\bar{z}) = B_\nu[T(0)]\tau_\nu(\bar{z}, 0) + \int_0^{\bar{z}} B_\nu[T(z)](d/dz)\tau_\nu(\bar{z}, z)\, dz \qquad (15a)$$

The surface flux $B_\nu[T(0)]$ is multiplied by the 'transmittance' of the atmosphere τ_ν (which has been integrated over all angles of flux direction from the surface, on the assumption of Lambert emission (proportional to the cosine of the angle from the normal to the surface), for the frequency ν and for the atmosphere between the surface $z = 0$ and the atmospheric level $z = \bar{z}$) to determine the surface contribution to $F_\nu\uparrow(\bar{z})$.

The second term in (15a) represents the contribution to the total upward flux from isotropic emission of infrared radiation by optically active gases below the level $z = \bar{z}$. Unlike the surface emission, which is nearly ideal blackbody radiation, the atmospheric emission as incorporated into the z derivative of τ_ν is highly frequency dependent because of the selective absorption of CO_2 or H_2O vibrational-rotational bands.

Similarly, the downward infrared radiation flux at $z = \bar{z}$ arriving from above $\bar{z}$ is given by

$$F_{\nu\downarrow}(\bar{z}) = \int_{\bar{z}}^{\infty} B_\nu[T(z)](d/dz)\tau_\nu(\bar{z}, z)\, dz \qquad (15b)$$

Thus the downward infrared flux results only from the atmospheric infrared emitters, since the incoming planetary infrared radiation from space is essentially zero.

The transmission of the atmosphere to infrared radiation for the frequency interval $(\nu, \nu + \Delta\nu)$ and between atmospheric altitudes $z = z_1$ and $z = z_2$ can be expressed in terms of the optical thickness of the atmosphere, $u(z_2, z_1)$, by

$$\tau_\nu(z_2, z_1) = (2/\Delta\nu) \int_\nu^{\nu+\Delta\nu} Ei_3[u(z_2, z_1)]\, d\nu \qquad (16a)$$

Here the monochromatic optical thickness is

$$u_\nu(z_2, z_1) = \int_{z_1}^{z_2} (\rho_1 K_{\nu_1} + \rho_2 K_{\nu_2} + \cdots)\, dz \qquad (16b)$$

and Ei_3 is the third-order exponential integral accounting for the integration of exponential attenuation over all zenith angles of radiative fluxes weighted according to the cosine of the angle from zenith direction. For physical interpretation it is useful to note that (16*a*) can in some cases be approximated by [*Kondratyev,* 1965, pp. 134–144]

$$\tau_\nu(z_2, z_1) \propto \exp\,[-\beta u_\nu(z_2, z_1)] \qquad (16c)$$

where $\beta = \sec \langle\zeta\rangle \simeq 1.66$, which indicates an average zenith angle $\langle\zeta\rangle$ of about 53°. However, for numerical calculations, fast routines are available to calculate the exponential integrals as rapidly as exponentials, and the former are more accurate.

One approach to computing transmission functions and infrared functions that has been used by *Rasool and Schneider* [1971] integrates over wavelength, using a smoothed set of 'generalized' absorption coefficients taken, for example, from the paper by *Elsasser and Culbertson* [1960]. Because of the wavelength integration, this method takes a considerable amount of computer time to provide the transmission through a given layer but offers the possibility of more accurately computing transmission, especially in regions where absorption bands of different constituents overlap (such as the 15-μm CO_2 band and the water vapor rotational bands).

Using this approach, *Schneider* [1972] computed the effect of variations in cloud top height and amount on the upward infrared radiative flux as follows. Clouds were taken into account simply by assuming that their boundary surfaces are black to infrared radiation; this is probably a good assumption for all but high thin cirrus clouds [*Manabe and Wetherald,* 1967; *Hunt,* 1973]. To evaluate $F_{\nu\uparrow}(\infty)$, the infrared flux at the wave number ν emitted to space by the earth-atmosphere system including the presence of clouds, Schneider assumed all clouds to lie in a thin layer at the same level. He broke (15*a*) into two terms and solved the following expression by numerical integration:

$$\begin{aligned} F_{\nu\uparrow}(\infty) = &\left\{ B_\nu[T(0)]\tau_\nu(\infty, 0) \right. \\ &\left. + \int_{\tau_\nu(\infty,0)}^{1} B_\nu[T(z)]\, d\tau_\nu(\infty, z) \right\}(1 - A_c) \\ &+ \left[B_\nu[T(z_c)]\tau_\nu(\infty, z_c) \right. \\ &\left. + \int_{\tau_\nu(\infty,z_c)}^{1} B_\nu[T(z)]\, d\tau_\nu(\infty, z) \right] A_c \end{aligned} \qquad (17)$$

where the subscript c refers to the z level of the cloud tops and A_c is the fraction of sky covered by the cloud layer. The first term inside the braces in (17) is the amount of radiation from the surface that escapes through the atmosphere directly to space, and the second term inside the braces is the atmospheric emission to space. The first term inside the brackets in (17) is the amount of radiation from cloud tops that escapes to space, and the second term inside the brackets is the atmospheric emission of infrared radiation to space that originates in the fraction of the atmosphere lying above the clouds. The total infrared flux escaping to space $F_{IR\uparrow}(\infty)$ was computed by a 240-step quadrature over wave numbers 10–2400 cm^{-1}:

$$F_{IR\uparrow}(\infty) = \int_{\nu=10}^{2400\ cm^{-1}} F_{\nu\uparrow}(\infty)\, d\nu \qquad (18)$$

The accuracy and the level of sensitivity being sought in the particular climate experiment dictate the details required in the infrared model. Often, band averaging (i.e., averaging the transmissivity with frequency and using a frequency-averaged Planckian function rather than integrating (15) over frequency) is of sufficient accuracy, since the uncertainties in heating rates from solar flux divergence calculations or convective adjustment procedures can easily exceed the difference between band averaging and wavelength integration in infrared calculations, as we shall discuss shortly. For band-averaged models where no wavelength integration is required the average transmittance $\tau(u_i)$ is usually interpolated from a semiempirically determined graph of τ versus u_i (as was done by *Manabe and Wetherald* [1967]) or is computed from formulas that fit the empirical data (as was done by *Sasamori* [1968]). The atmosphere is divided into many layers in the vertical direction. Thus τ_i is determined from the amount u_i by applying at each level a correction to the increment of path length depending on the temperature and pressure dependence of band strengths, individual line width, and frequency distribution.

For certain types of sensitivity experiments where radiative fluxes related to the varied parameter are not strongly wavelength dependent, as is the case for changes in cloudiness, band-averaged models may suffice; whereas for other experiments (for example, situations where overlap between absorption spectra of different constituents can seriously affect the sensitivity response in the model) some wavelength integrations (at least in the regions of overlap) may be required. For instance, Manabe and Wetherald computed a +2.36°K increase in surface temperature from a doubling of atmospheric CO_2 content, using a band-averaged calculation, whereas *Manabe* [1971] reported only a +1.9°K temperature increase, using a model identical to his former one except for the subdivision into many limited spectral intervals, where the CO_2–water vapor overlap region around 15 μm is divided into four subintervals.

Absorption of solar energy by the entire earth-atmosphere system is most simply specified in terms of the albedo of that system. Usually, separate albedos for the cloudless part of the earth-atmosphere system (about 0.14 according to *London and Sasamori* [1971]) and the cloud-covered part (about 0.50 on a global average) are used. Applying (17), the lapse rate being assumed constant, Schneider showed (as is inferred from Figure 11) that increasing the cloud amount (at fixed cloud top height and albedo) by about 8% would lower the global mean surface temperature by 2°K, whereas raising the level of the cloud top height by about ½ km (at constant cloud amount and albedo) would produce exactly the opposite effect on surface temperature.

The study showed that it is not possible to generalize globally averaged results to yield information on the local (in latitude and time) effect of variations in cloudiness, since the effect of changes in cloudiness on surface net heating depends upon the local values of the cloud amounts, heights, and

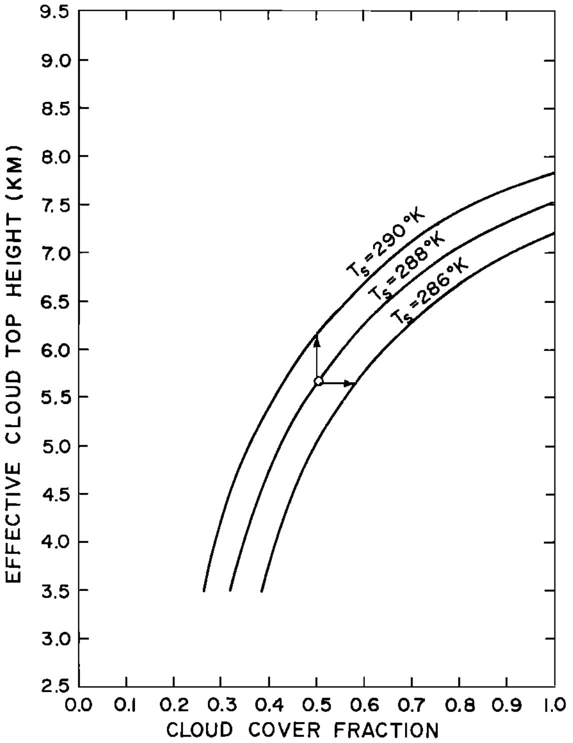

Fig. 11. Effective cloud top height and cloud cover fraction giving the indicated equilibrium surface temperatures according to *Schneider* [1972].

albedos, the albedo of the surface, the average solar zenith angle, and the local vertical distribution of temperature and optically active constituents. The radiative calculations do suggest that even the direction of possible cloud feedback on surface temperature is far from obvious (see section B2*d*).

In models that determine the vertical distribution of solar flux divergence [e.g., *Manabe and Strickler,* 1964; *Sasamori et al.,* 1972] the absorption of sunlight by water vapor, ozone, and clouds and the scattering by clouds and air molecules have generally been included. For example, the model of *Manabe and Strickler* [1964], which has been the basis for the radiation calculations in subsequent GFDL atmosphere GCM's, is based on the following assumptions:

1. Solar radiation becomes completely diffused after it enters a cloud or after it is reflected by a cloud.
2. All the reflection of insolation by a cloud takes place at the upper surface of the cloud.
3. The effective optical thickness for diffuse radiation is assumed to be 1.66 times that for direct solar radiation (equivalent to approximation (16*c*)).
4. The surface albedo is 0.102 except in a few special areas assumed to be covered by ice and snow.
5. Rayleigh scattering of solar flux by air molecules reduces the solar constant Sc by 7% to Sc'; i.e., $Sc' = Sc(1 - 0.07)$, whereas the influence of Rayleigh scattering on solar absorption by the atmosphere is neglected.

Figure 12 illustrates the processes involved in computing the absorption of solar radiation. The total daily mean downward flux of solar radiation at level z is calculated by

$$S_z = (1 - C)\cdot(Sc' - A_z)\cdot\cos\zeta\cdot r + C\cdot[1 - (\alpha + \beta)] \cdot(Sc' - OA_z)\cdot\cos\zeta\cdot r \qquad (19)$$

where α and β are the cloud reflectivity and the absorptivity of clouds due to droplets, respectively; C is the cloud amount; and ζ is the effective mean zenith angle of the sun, defined by

$$\cos\zeta = \left[\int \cos\zeta(t)\,dt\right]\Big/\left(\int dt\right) \qquad (20)$$

in which the integrations are carried out only for hours of daylight. Also, r is defined as the total daylight hours divided by 24 hours. The albedos and the absorptivities of clouds due to droplets are given in Table 3. Terms A_z and OA_z are the gaseous absorption of solar radiation from the top of the atmosphere to level z by a clear atmosphere and by an overcast atmosphere, respectively, computed from semiempirical band-averaged absorptivity curves for water vapor, carbon dioxide, and ozone.

The total upward reflected radiation at level z is computed as the sum of radiation reflected by clouds and that reflected by the earth's surface. The effect of the depletion of reflected radiation by clouds is neglected. The total absorption of solar radiation was computed by adding the absorption of downward solar radiation and that of reflected radiation.

Recently, *Lacis and Hansen* [1974] have given an improved parameterization for the computation of vertical solar radiative fluxes.

b. Convective adjustment. Equation (13*b*) defines a radiative temperature profile $T_r(z)$, determined solely from the vertical divergence of the net radiative fluxes. Figure 13 [after *Manabe and Wetherald,* 1967] shows that for clear sky conditions and with either a fixed distribution of relative humidity (the solid line on Figure 13) or a fixed distribution of absolute humidity (the dashed line on Figure 13), computation of globally averaged vertical radiative temperature profiles yields

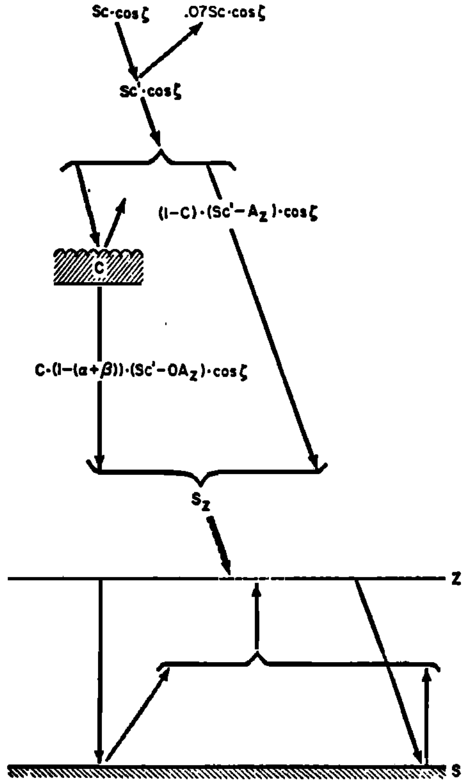

Fig. 12. Vertical distribution of the flux of solar radiation in a model atmosphere with clouds [*Manabe and Strickler,* 1964].

TABLE 3. Albedos and Absorptivities of Clouds Due to Droplets

Cloud Level	Albedo	Absorptivity of Clouds Due to Droplets
High	0.21	0.005
Middle	0.48	0.02
Low	0.69	0.035

very high surface temperatures, decreasing extremely rapidly with altitude. Manabe and Wetherald argue that fixed relative humidity is the more realistic condition for the earth's atmosphere, but both assumptions give radiative temperature profiles $T_r(z)$ with large lapse rates in the lower troposphere, considerably in excess of the mean value given by the U.S. Standard Atmosphere (1962) lapse rate $\gamma_c = 6.5°K/km$. It is usually argued that the observed lapse rate γ_c is less than $-\partial T_r/\partial z$ in the lower troposphere, having a nearly constant value up to about 10 km because the radiative equilibrium profiles are modified by free and forced vertical (moist and dry) convection and the vertical heat transport due to large-scale eddies. Radiative equilibrium profiles are unstable to vertical (moist and dry) convection. Thus the tendency of radiative heating alone to produce large lapse rates directly above the earth's surface, as shown by the solid and dashed curves on Figure 13, is offset by a rapid vertical transfer of heat. The procedure in the model that mimics the vertical heat flux due to convective motions and latent heat release is called convective adjustment. It is applied whenever the critical lapse rate γ_c is exceeded in the time evolution of the numerical calculation, i.e., (13*a*), by adjusting the temperature between vertical layers to the critical lapse rate γ_c and by changing the temperature with time according to the integrated rate of heat addition. Radiative-convective adjustment has also been used for steady state astrophysical and geophysical calculations with the additional condition that the temperature across the interface between the convective and the radiative equilibrium regions remain continuous. *Gierasch and Goody* [1968] have shown that in the course of a time-dependent calculation, if continuity of temperature across this interface is satisfied at one time, it will be satisfied for all later times.

A critical lapse rate of 6.5°K/km, as assumed by Manabe and Wetherald in their calculations, is actually less (i.e., more stable) than that which would be critical for dry convection, since the dry adiabatic lapse rate $\gamma_d = g/C_p$ has a value of 9.8°K/km for the earth's atmosphere. On the other hand, γ_c is often supercritical for moist convection, since the moist adiabatic lapse rate

$$\gamma_w = \gamma_d \left[\frac{1 + a[q_s(T)/T]}{1 + b[q_s(T)/T^2]} \right] \tag{21}$$

is usually less than 6.5°K/km over a large range of temperatures and pressures found in the lower troposphere (as discussed in any basic meteorology text) and shown on Figure 5. In (21), q_s is the saturation mixing ratio of water vapor, and a and b are constants; γ_w is always less than γ_d but approaches γ_d as pressure increases or temperature decreases. Thus the 6.5°K/km U.S. Standard Atmosphere lapse rate γ_c is intermediate between γ_d and γ_w throughout much of the troposphere, particularly in the tropics (i.e., γ_w can be greater than γ_c only for very cold or very dry conditions).

The fact that $\gamma_w < \gamma_c < \gamma_d$ for much of the atmosphere suggested to GCM modelers [e.g., *Manabe et al.*, 1965; *Washington and Kasahara*, 1970] that the convective adjustment procedures in their GCM's could be split so that supermoist adiabatic lapse rates would be adjusted to γ_w at grid points where moist convection was assumed to occur, whereas dry adiabatic adjustment would be applied at other grid points that also had a superdry adiabatic lapse rate. Manabe et al. invoked moist adjustment if saturation occurred at the same time that the convective adjustment was used, whereas Kasahara and Washington used the following simple procedure: if $-\partial T_r/\partial z > \Gamma_c$, then $-\partial T/\partial z$ is set equal to Γ_c, where

$$\begin{aligned} \Gamma_c &= \gamma_d \qquad w < 0 \\ \Gamma_c &= \gamma_w \qquad w > 0 \end{aligned} \tag{22}$$

and where w is the vertical velocity.

Since q_s depends upon temperature and since precipitation follows the processes of condensation and the release of latent heat, which in turn affects temperature and thus q_s, simultaneous inclusion of convective adjustment and precipitation processes may cause subsequent supercritical lapse rates and/or supersaturated layers. Thus *Manabe et al.* [1965] adopted an iteration procedure to insure that the latter do not occur.

The classical convective adjustment procedure will not produce the well-mixed atmospheric boundary layers (i.e., where $\partial T/\partial z = -\gamma_d$, or equivalently, $\partial\theta/\partial z = 0$, where θ is potential temperature) seen on Figure 7, unless heating at the surface creates an unstable atmospheric temperature gradient. However, shallow, well-mixed layers do occur even in the absence of surface heating owing to turbulent mixing in the atmospheric boundary layer, as has been described in section C3*d* for the oceans. This atmospheric well-mixed layer does not have quite the same significance for climate as the oceanic mixed layer. Like the oceanic boundary layer, it is an important factor in determining the magnitude of vertical fluxes away from the surface [*Deardorff*, 1972], as discussed in sec-

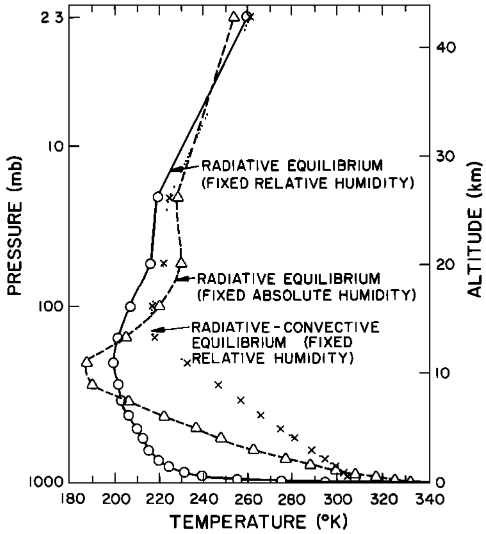

Fig. 13. Convective adjustment. The solid line shows the radiative equilibrium of the clear atmosphere with the given distribution of relative humidity, the dashed line shows the radiative equilibrium of the clear atmosphere with the given distribution of absolute humidity, and the dotted line shows the radiative-convective equilibrium of the clear atmosphere with the given distribution of relative humidity [*Manabe and Wetherald*, 1967].

tion C3*c*. However, the oceanic mixed layer governs not only interfacial vertical transports but also the thermal relaxation time of the surface layers, because of the large heat capacity of the oceanic mixed layer; whereas the atmospheric mixed layer has relatively negligible thermal capacity.

The real atmosphere is not usually saturated in the lowest atmospheric layers, especially in the tropics [e.g., *Holton*, 1972, Figure 12.1; *Ooyama*, 1969, Figure 2]. Thus cumulus convection may not necessarily be initiated even if the actual lower atmospheric lapse rate is slightly greater than the moist adiabatic rate and thus is unstable to moist convection—unless the lower layers are also at or near saturation. The presence of some other mechanisms (such as low-level convergence) may be required to lift moist, convectively unstable (but unsaturated) surface air to a condensation level before convective precipitation is possible. The potentially key role of tropical convection as an energy source for the general circulation has generated much interest and has led to the initiation of a large-scale field program: the Garp (global atmospheric research program) Atlantic tropical experiment (Gate), to be carried out in mid-1974.

Schemes for improved parameterization of moist convection in GCM's are under development, including the cumulus parameterizations of *Arakawa and Schubert* [1974] or *Arakawa et al.* [1968], which are also described by *Haltiner* [1971]. Basically, Arakawa considers several types of convection: low level, middle level, and penetrating, which occur according to the relative buoyancy of the lower, middle, and upper levels of the troposphere. Adjustment of environmental parameters such as temperature or humidity as a result of moist convection depends upon which type of convection is present. Recently, a version of Arakawa's scheme has been adopted to the nine-level GISS model and described by *Somerville et al.* [1974].

2. *Energy Balance Models*

Earlier (in section C2*b*) we defined energy balance models as those that emphasize the calculation of surface temperature in terms of a balance between incoming solar and outgoing infrared radiation. The horizontally averaged one-dimensional vertical models discussed above could, strictly speaking, be considered a subset of energy balance models, but our discussion in this section will refer specifically to energy balance models whose vertical structure is treated with much less detail than their horizontal structure. In particular, the early models of *Budyko* [1969] and *Sellers* [1969] will be reviewed in some depth because their rather startling predictions served to focus interest on climate modeling and their features illustrate many aspects common to energy balance models.

a. Function of latitude only. In view of the past record of large fluctuations in climate, one of the major topics of concern about the possible impact of man's activities has been the question of the stability of the global climate to small perturbations of any sort. This question led *Budyko* [1968] and *Sellers* [1969] (independently) to develop models of the earth's climate based on the equations of the zonally averaged heat balance of the earth-atmosphere system and for the purpose of testing the sensitivity of the equilibrium state of the climate to changes in solar radiative heating. One important parametric relationship common to their models is the coupling between the planetary albedo and the temperature of the atmosphere near the earth's surface. This coupling provides a strong positive feedback between ice cover and temperature as discussed in section B2*c*. (Some connections between this feedback and that of baroclinic eddy processes are discussed below in section D3*d*.) Thus Budyko and Sellers found as a consequence of the positive feedback in their models that changes in solar heating of the order of 1% could result in extensive changes to the equilibrium climate (e.g., a near melting of the ice caps or a significant expansion of glaciation). These results, although they are based on extremely simplified parameterizations of most atmospheric transport processes, are nevertheless sufficiently compelling to indicate the need for further study of climatic stability in order to determine just how large a perturbation to the environment would be required, when the various climatic feedback processes are allowed for, to alter our climatic regime significantly [*Budyko*, 1969, 1972].

More recently, *Schneider and Gal-Chen* [1973] have explored the sensitivity of time-dependent models of the Budyko-Sellers type to changes in initial as well as external conditions. Since the Budyko-Sellers models are formulated independently of time, the solutions are automatically equilibrium 'climatic' statistics. The stability of their models' equilibrium climates was tested only by variations in externally prescribed parameters—such as the solar input, amount of atmospheric turbidity, boundary conditions, etc. However, Schneider and Gal-Chen formulated the problem as an initial value problem in which the basic equations (which were similar to those of Budyko and Sellers) were integrated from a given initial state until a final asymptotic equilibrium state (i.e., climate) was obtained. It is instructive to review this more recent work, since the results combined many of those of Budyko and Sellers and showed how they extended to models evolving in time.

A time-dependent version of the zonally averaged, vertically integrated equation of the heat balance of the earth-atmosphere system is written as

$$C\, \partial T/\partial t = R + \mathrm{div}\,(\mathbf{F}) \tag{23}$$

where C is the thermal inertia coefficient for a zonal column through the land-ocean-atmosphere system (its value was chosen by following the method of *Sellers* [1965, p. 242]) based on a 25-m mixed layer in the oceans and a linearly varying temperature profile down to 0°C at a depth of 125 m. The value of C could also be interpreted as a scaling factor for the time scale t provided that C were changed by the same factor at all latitudes ϕ. In these computations, C $(=C(\phi))$ is made proportional to the fraction of the area of a latitude circle covered by oceans. Also, T is the zonally averaged 1000-mbar temperature, and R is the radiation balance term for a zonal column through the earth-atmosphere system:

$$R = Q(1 - \alpha) - \Delta I \tag{24}$$

where $Q(\phi)$ is the yearly averaged, zonally averaged value of solar energy input at latitude ϕ, α is the planetary albedo at latitude ϕ, and ΔI is the net infrared cooling to space assumed to have the temperature dependence

$$\Delta I = c(\phi)\sigma T^4[1 - m \tanh (19T^6 \times 10^{-16})] \tag{25}$$

where the opacity factor m is 0.5 and $c(\phi)$ is a consistency factor to be described later. With $c(\phi) = 1$, (25) is identical to the infrared flux formulation of *Sellers* [1969]. Finally, div (**F**) is the net energy flux transport for latitude ϕ and is parameterized differently for the various models used.

Four parameterizations were used by Schneider and Gal-Chen. The first one (S) is written explicitly as a divergence of energy flux terms:

$$\operatorname{div}(\mathbf{F}) = \frac{1}{\cos\phi}\frac{\partial}{\partial y}\cos\phi(F_O + F_A + F_q)$$

where $y = a\phi$, a being the radius of earth and ϕ being the latitude belt; F_O is the sensible heat transport due to ocean currents; F_A is the sensible heat transport due to atmospheric motion; and F_q is the transport of latent heat energy. The energy flux terms in formulation S are parameterized by

$$F_O = -K_O\, \partial T/\partial y$$

$$F_A = -K_A\, \partial T/\partial y + \langle v\rangle T$$

$$F_q = -K_q\, \partial q(T)/\partial y + \langle v\rangle q(T)$$

where $q(T)$ is proportional to the water vapor mixing ratio and $\langle v\rangle$ is a vertically integrated mean meridional velocity and is parameterized as a function of latitudinal temperature gradient. The K are effectively eddy diffusion coefficients. The assumed albedo is:

$$\begin{aligned} \alpha &= b(\phi) - 0.009T \qquad & T < 283°K \\ \alpha &= b(\phi) - 0.009 \times 283 \qquad & T > 283°K \\ 0.25 &\leq \alpha \leq 0.85 & \text{all } T \end{aligned} \qquad (26)$$

If (26) gives $\alpha(T) > 0.85$, α is set equal to 0.85, and if (26) gives $\alpha < 0.25$, α is set equal to 0.25. Formulation S is based on work of *Sellers* [1969], from which further details can be obtained.

The second formulation (SV) is similar to formulation S except that $\langle v\rangle \equiv 0$.

The third formulation (B) is div (F) = $-\beta(T - \langle T_p\rangle)$, where β is an empirically determined coefficient given by *Budyko* [1969] as $\beta = 0.235$ (kcal/cm²/month/°K) and $\langle T_p\rangle$ is the mean (global average) planetary temperature (at 1000 mbar). Formulation B is based on the dynamic parameterization for mean heat transfer by *Budyko* [1969].

The fourth formulation (FG) is similar to formulation S except that the albedo is assumed to be a function of temperature everywhere:

$$\alpha = 0.4860 - 0.0092(T - 273) \qquad (27)$$
$$0.25 \leq \alpha \leq 0.85 \text{ all } T$$

This formulation is based on that of *Faegre* [1972].

In order to evaluate the eddy diffusion coefficients, observed energy fluxes F_O, F_A, and F_q [*Sellers*, 1969, Figure 2] and $\langle v\rangle$ and T are all taken from *Sellers* [1969]. From these empirical data and, for example, parameterization S the various eddy coefficients are computed. Since formulation SV neglects 'advection' (i.e., $\langle v\rangle \equiv 0$), it requires several of the eddy coefficients in tropical latitudes to be negative.

When the empirical values described above (i.e., those for F_O, F_A, F_q, $\langle v\rangle$, $T(\phi)$, $q(T)$, $\alpha(\phi)$, and ΔI) are used and when (23) is written in finite difference form, the right-hand side of (23) does not identically vanish. That is, $\partial T/\partial t \not\equiv 0$ at $t = 0$, and the resulting asymptotic equilibrium climatic state (i.e., the $T(\phi)$ that satisfies $\partial T/\partial t = 0$ in (23)) is not identical to the initially prescribed temperature distribution. Therefore appropriate parameters should be adjusted so that the consistency constraint

$$R_e = -\operatorname{div}(\mathbf{F}) \qquad t = 0 \qquad (28)$$

is imposed in finite difference form at the initial time and for the control climate case.

Despite the inherent uncertainties in the absolute accuracy of the empirical data (in some cases, up to 100%) the initial control climate should be closely balanced (i.e., (28) should be satisfied) because the intention is to test the sensitivity of the model's climates to variations as small as 1% in some parameters. Consistency could be achieved by adjusting each of the energy flux terms F_O, F_A, and F_q (see formulation S) or alternatively by adjusting the radiation balance term R or the initial temperature distribution. However, Schneider and Gal-Chen found by numerical experimentation that the final conclusions were not significantly dependent on the particular choice of term that was altered, provided that the alteration was essentially within the range of observational uncertainty.

Thus in cases S, SV, and FG it was decided to achieve consistency through adjusting the factor $c(\phi)$ in (25). This factor $c(\phi)$ was found to be in the range

$$0.97 \leq c(\phi) \leq 1.03 \qquad (29a)$$

for S and SV and in the range

$$0.92 \leq c(\phi) \leq 1.04 \qquad (29b)$$

for FG. In view of the uncertainties implicit in the use of a single-level temperature formula such as (25) for the evaluation of outgoing planetary infrared radiation and in view of the fact that zonal asymmetries in cloudiness, which are not specifically accounted for in (25), could significantly alter the latitudinal dependence of ΔI on T, Schneider and Gal-Chen felt that the range shown in (29) was acceptable, i.e., within observational uncertainties. In case B, $c(\phi) \equiv 1$, and consistency was achieved through adjustment of the albedo latitude constant $b(\phi)$ (see (26)).

The need to make both the initial surface temperature and the initial surface energy balance (which, of course, depends upon the surface temperature) consistent with observed values for these distributions is also important for other types of climate models. For example, in order to couple an atmospheric GCM (for which sea surface temperature is presently specified) to an ocean surface temperature prediction model, great care must be used in insuring that the combined models' surface energy balance in the mean (i.e., G in (3)) nearly vanishes with conditions appropriate to the present climate. Otherwise, a large surface temperature change would result. For instance, an absolute imbalance in the surface energy balance of, say, 10% would give an error in surface temperature of the order of 10°C. Such an initial imbalance might, for example, be corrected by adjusting cloud albedos or amounts, surface drag laws (e.g., (5)), or the surface hydrology parameterization. Of course, such adjustments should be made so that the adjusted parameter remains within observational uncertainties, but this still leaves a great deal of latitude for justifiable 'tuning.' A model not so tuned may yield significantly erroneous results, since many processes are nonlinearly coupled to the surface temperature (such as the distribution of ice and snow or evaporation).

In one of the parameterizations for energy flux used by Schneider and Gal-Chen (S) and in that of Sellers the same value of $\langle v\rangle$ is used for the meridional advection of sensible heat by the atmosphere as is used for the advection of latent heat by the atmosphere. It may, however, be difficult to associate this term with some weighted average of the observed meridional circulation. In view of the rapid decrease of the water mixing ratio with altitude the mean cell flux of latent heat depends primarily on the meridional circulation near the ground, whereas the mean cell flux of sensible heat depends on the flow throughout the depth of the atmosphere [e.g., *Oort and Rasmussen*, 1971, Figures A2, A4, and A6]. Also, as was pointed out by *Robinson* [1971], the $\langle v \rangle$ used by *Sellers* [1969] does not obey the mass continuity equation. Sellers himself has already pointed out several further difficulties with the $\langle v \rangle$ parameterization and has justified its use primarily as a means of eliminating the need for negative diffusivities. Nevertheless, as is now discussed, the asymptotic equilibrium states of the models approached from different given initial conditions appear much less sensitive to the details of the div (F) formulation than to the parameterization of the coupling between temperature and albedo.

The climate in the steady state models of Sellers and Budyko was found to change drastically as a result of small changes in the value of the solar constant. Verification of the time-dependent version (hereafter referred to as the SG model) of these models requires that the SG model be capable of recovering the results of Sellers and Budyko. That is, the asymptotic equilibrium climate from the SG model should exhibit a sensitivity to changes in the value of the solar constant similar to that exhibited by the previous models, provided that the SG initial state is close to the present-day state and is essentially balanced for present-day conditions. Such agreement of the SG asymptotic state with the previous steady state solutions is found, as is indicated in Figure 14, for the Sellers flux parameterization (S). Curve *a* on Figure 14 shows the asymptotic equilibrium climate (given by about 10 years of 2½-day time steps), having a mean 1000-mbar temperature of 287.06°K; it will henceforth be referred to as the 'control climate.' The average value of the control temperature distribution differed slightly from the average value of the initial temperature distribution of 287.30°K owing to roundoff errors from the computer used. Curves *b* and *c* result from a 1% increase or decrease, respectively, in the value of the solar constant. A 1% increase in solar input increases the mean global temperature to 289.51°K (shown by curve *b* on Figure 14), which is significantly larger than the ≈1.5°K increase that would be obtained in the absence of positive feedback between the albedo and the temperature. On the other hand, the 1% decrease in solar constant shown in curve *c* yields a mean global temperature of 281.22°K, a factor of 4 amplification of the decrease in temperature due to the positive feedback between planetary albedo and surface temperature. With a decrease in solar constant of more than 1.6% a planetary mean temperature of 174.65°K was obtained, corresponding to an entirely ice-covered earth.

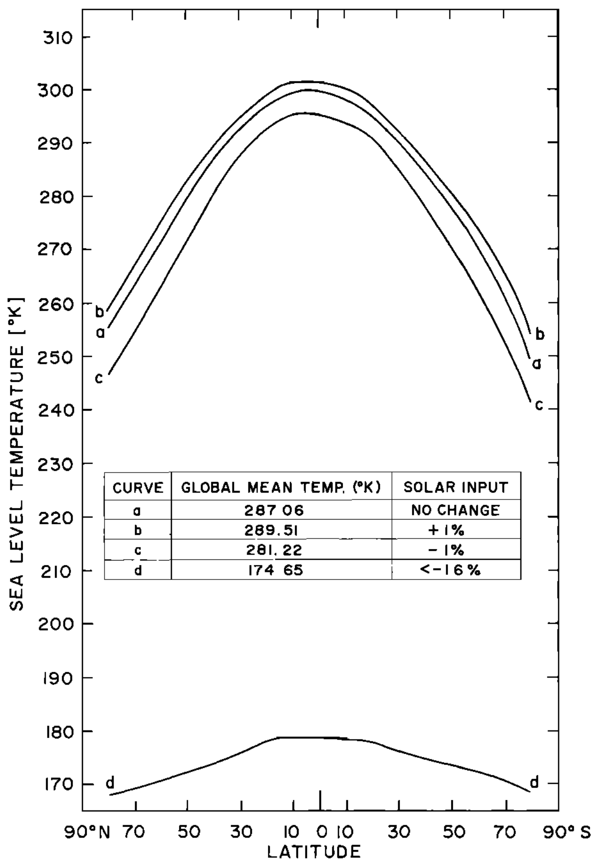

Fig. 14. Steady state temperature distributions for various values of solar input and parameterization S [*Schneider and Gal-Chen*, 1973], showing the calculated extreme sensitivity of surface temperature to small variations in solar input.

Calculations for a 1% decrease in solar constant and Sellers-type energy fluxes with (S) or without (SV) mean meridional motion give quite similar results both in the global average amount of temperature decrease and in the latitudinal distribution of the temperature decrease. It is important to point out that ice feedback implies an increased equator-to-pole temperature gradient with decreased solar input, hence probably increasing eddy horizontal and vertical fluxes (as is discussed in section D3*d*). The results for the Budyko parameterization are qualitatively similar to those for formulations S and SV for a global average. However, the latitudinal distribution of temperature decrease differs significantly from that of Sellers. Thus although the results for the global mean temperature as well as the conclusions from stability experiments (as discussed below) using these models are similar, the particular form of the horizontal energy flux parameterization does appear to be significant for determining the latitudinal distribution of temperature.

In the stability experiments a perturbation was added to the initial temperature distribution, whereas all other parameters in the SG models were held fixed. It was found in both the Sellers and the Budyko parameterization cases, as summarized in Table 4, that the steady state control climate, i.e., T = 287.06°K in the Sellers case (S) and T = 287.09°K in the Budyko case (B), could be recovered exactly, provided that the perturbation added to the initial temperature was greater than 0. Perturbations of +2°K and +16°K were tried, and in both cases (the latter of which implies a melting of the polar ice caps), energy balance requirements alone produced a return to the control climate; i.e., unique, fully stable, transitive (see sections A3 and A4) control climates with mean temperatures of 287.06°K and 287.09°K for the Sellers and Budyko dynamic formulations, respectively, were recovered. However, for an initial temperature decrease of a magnitude

TABLE 4. Asymptotic Steady State Global Average Temperatures Resulting From Various Perturbations to the Initial Temperature Distribution for Two Parameterizations

Initial Temperature Perturbation	Mean Planetary Temperature, °K	
	Sellers (S)	Budyko (B)
0	287.06	287.09
$0 < P < 16$ for all ϕ	287.06	287.09
$-18.3 < P \leq 0$ for all ϕ	286.67	286.87
$P < -18.3$ for all ϕ	175.58	175.44
-22 for $\|\phi\| \geq 20°$ 0 for $\|\phi\| < 20°$	286.67	286.87

not exceeding 18.3°K (i.e., −18.3°K < perturbation < 0°K) the equilibrium climate was slightly intransitive; i.e., the final equilibrium climate derived was 286.67°K and 286.87°K for the Sellers and Budyko parameterizations, respectively. These values are a few tenths of a degree lower than the control climates but are still very close to the present-day control equilibrium statistics. (This slight intransitivity has recently been attributed to computer rounding errors (as mentioned above for the control cases).) The final latitudinal distributions of temperature were also always very similar, so that particular climatic states can be represented by the global average values. In summary, the models show the asymptotic equilibrium climate to be extremely insensitive to changes in initial conditions. A 17°K perturbation in initial temperature is an incredibly large decrease, yet the asymptotic solution was able to return essentially to the present equilibrium state.

Most interestingly, however, initial temperature decreases greater in magnitude than about −18.3°K in both the Sellers and the Budyko parameterization cases lead to an ice-covered earth solution similar to curve *d* of Figure 14 (which was derived from a 1.6% decrease in solar constant, an external parameter). Of course, in these initial temperature perturbation experiments the solar constant is fixed at the control value. Thus it takes an extremely large negative initial temperature perturbation to lead into the highly intransitive alternative ice-covered earth regime of these models. The required perturbation of −18°K is interpreted in terms of the control temperature distribution given by curve *a* of Figure 14. It is apparent that the surface temperature is about 300°K in the equatorial latitudes. Equation (26) for the albedo formulation shows that only mean zonal temperatures less than 283°K yield a temperature-dependent albedo. The difference between 300 and 18 is 282, and thus it appears that the insensitivity of the model to decreases in initial temperature conditions of less than 18°K follows from an unchanged albedo in equatorial latitudes.

The implication of these experiments is that even when the temperate and polar latitudes are ice covered by virtue of the large negative initial temperature perturbation but the solar constant and tropical albedo are unaltered, there is still enough energy available to the tropics to provide heat transport poleward to 'melt the ice' and to restore the equilibrium climate to near the present-day value. However, if the equatorial temperature is decreased sufficiently (i.e., by more than 18.3°K) by the initial temperature perturbation so as to lower the tropical surface temperature into the range of the positive feedback temperature-albedo coupling implicit in (26), then climatic instability is triggered, and the ice-covered earth regime results. Thus the ice-covered earth solutions of both the *Budyko* [1969] and the *Sellers* [1969] models that were obtained by a decrease in the solar constant of more than a few percent appear to result from a reduction of the solar constant everywhere and in particular from a reduction of the energy available to the model tropics. The model tropics were consequently no longer able to supply the energy export necessary to prevent the positive feedback albedo-temperature coupling in the model temperate and polar latitudes (where $T < 283°K$) from causing a runaway ice age. This hypothesis was tested further by application of a very large negative perturbation in initial temperature, i.e., −22°K, to all latitudes poleward of 20°. The initial temperatures at the equator and at latitude zones 10°N and 10°S were unchanged. In both the Sellers and the Budyko parameterization cases (Table 4) the result was a return to near the present-day climate. Thus an extreme initial ice age in temperate and polar latitudes does not persist and is unable to cause a runaway ice-covered earth if the present level of energy input to the tropics is maintained. This finding must in part be due to the negative feedback in the parameterizations of both Budyko and Sellers, where increased poleward energy transport occurs with an increase in the equator-to-pole temperature contrast. This feedback would be even stronger if the *Stone* [1972*a*] eddy parameterization were used (cf. section D3*d*).

Finally, we consider the albedo formulation of *Faegre* [1972] (FG). A great deal of difficulty was encountered in preventing the ice-covered earth asymptotic solution with Faegre's albedo formulation (27). Any negative perturbation to the initial temperature distribution gave an ice-covered earth as the asymptotic climate. However, for a perturbation of +0.152°K Schneider and Gal-Chen obtained approximately the present equilibrium climate, a planetary temperature of 287.45°K. For a perturbation in initial temperature of greater than +0.315°K the climate always returned to an equilibrium steady state solution corresponding to a global mean temperature of 301.95°K.

For different initial perturbations between +0.152°K and +0.315°K, correspondingly different equilibrium temperatures, all being quite close to the present-day equilibrium solution (T = 287.45°K), were obtained (resulting again from computer rounding errors). A +2°K perturbation with unaltered solar constant results in an extremely warm equilibrium global average temperature of 301.95°K. However, when in addition a 1% decrease in solar constant was assumed, an ice-covered earth equilibrium solution was obtained. Thus the Faegre albedo formulation is very sensitive to negative perturbations in both solar constant and initial conditions. The interpretation of these results is that the albedo formulation chosen by Faegre (equation (27)) includes a temperature-albedo coupling for all surface temperatures and at all latitude zones, and thus a slight negative perturbation in either initial temperature or solar constant can cause a decrease in the energy input to the tropics, thereby enhancing the albedo-temperature positive feedback mechanism at all latitudes. An albedo formulation such as Sellers' (equation (26)) would appear to be more realistic in applying the albedo-temperature coupling only with a zonal belt ground temperature less than some value, e.g., 283°K. It seems unlikely that the albedo of the tropics would be significantly affected by a temperature change of a few degrees in that region, since the amount of snow and ice cover is negligible in those zones—unless the amount of cloud cover were much more strongly coupled to the surface temperature than is generally believed (see discussion in section B2*d*).

In summary, the experiments with energy balance models show that the stability of the earth's climate may be quite dependent on the functional relationship between the albedo and the surface temperature. The model experiments also suggest that if the energy input to the tropical latitudes is left unchanged, large changes in the temperature (and hence the albedo) of middle and polar latitudes can eventually be ameliorated by exportation of sufficient energy from the tropics. Ice sheets have possibly never proceeded into the subtropics because the energy input to the tropics has always been sufficient to sustain poleward energy fluxes adequate for preventing the further expansion of the ice sheets and in fact eventually leading to their recession. Any such conclusion from the model must be viewed with caution, however, because the albedo-temperature formulation used in these models does not explicitly include coupling to the hydrological cycle. It must be borne in mind that the results obtained with such simplified empirical parameterizations cannot assuredly be close to reality. Nevertheless, the models do satisfy energy balance of the earth-atmosphere system, and in the long term it is energy balance that must determine the surface temperature. Thus these models are at the very least a valuable 'educational toy,' as they have been described by *Robinson* [1971], and their results suggest that climatic stability and transitivity experiments should be attempted with much more sophisticated models (atmospheric, oceanic, and joint atmosphere-ocean) that explicitly include detailed hydrological cycle and dynamic calculations. (Recent results with the GFDL GCM do indeed show similar effects of ice albedo feedback, as described in section D3*d*.)

b. Function of latitude and longitude. In section C2*b* we discussed briefly several energy balance modeling studies, including the study by *Sellers* [1973]. This study was an extension of Sellers' earlier model (see preceding section) to include some longitudinal dependence, and it differed in two fundamental respects: (1) it was time dependent, and (2) it included a seasonal cycle. Since the earth's climate undergoes significant and grossly predictable seasonal changes, a significant test of the veracity of a climate model is its ability to forecast the seasonal cycle. Therefore we will further examine the results of *Sellers* [1973].

In this study Sellers worked with the thermodynamical energy equation for the earth-atmosphere system applied separately to the land- and water-covered portions of each 10° latitude belt of an earth with a single large continent extending from pole to pole.

The model was formulated in terms of the cleverly conceived idealized continent-ocean system shown in Figure 15. Except for latitudes between 40° and 70°S the fraction of each 10° latitude belt occupied by land is equal to the actual current value. Furthermore, the land masses have been offset in relation to one another in order to give a meridional transport across each 10° latitude circle from water to land, water to water, land to land, and land to water, similar to that which actually exists. This offsetting accounts for the asymmetry in Figure 15. The dashed meridional lines over the continent separate the land masses of the eastern (to the right) and western (to the left) hemispheres. In order to simplify the calculations without drastically modifying the results, Sellers climatic 'Gonwanaland' has introduced a narrow bridge between 40° and 70°S connecting South America and Antarctica.

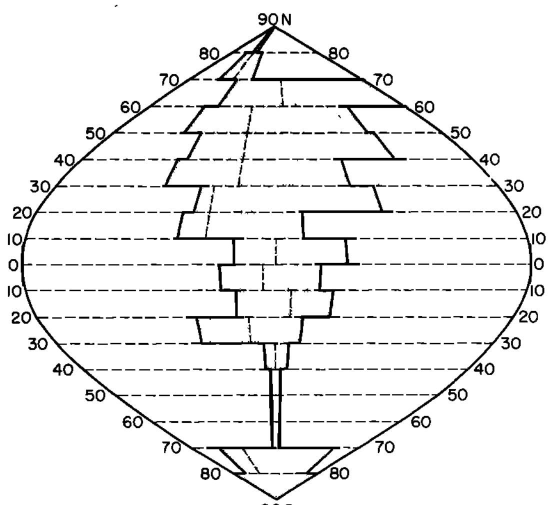

Fig. 15. Land and ocean distribution for the model of *Sellers* [1973].

Each term of the energy equation was written in terms of surface variables; e.g., the equation for the vertical temperature profile $T(p)$ was written in terms of the surface temperature T_s and the vertical temperature gradient $\partial\langle T\rangle/\partial p$ (p is pressure):

$$T(p) = T_s - (p_s - p)\ \partial\langle T\rangle/\partial p \tag{30}$$

where $\partial\langle T\rangle/\partial p$ is specified (as 0.12°K mbar^{-1}) and the angle brackets refer to a 1-month time average.

Eddy diffusivities were used to parameterize the poleward transport of heat by ocean currents and atmospheric standing and transient eddies. Unlike the earlier *Sellers* [1969] model, in which the eddy coefficients were fixed, this *Sellers* [1973] model assumed that the values of the atmospheric eddy coefficients for sensible heat K_H and water vapor K_v (assumed to be equal) are proportional to the first power of the north-south temperature gradient:

$$K_H = K_v = 0.25|\Delta T| \times 10^{10}\ \mathrm{cm^2\ s^{-1}} \tag{31a}$$

where ΔT is the temperature gradient between successive 10° latitude belts.

An eddy diffusivity K_w for heat transfer by ocean currents was specified as

$$K_w = 1.7 \times 10^9 \times (1 - A_I)A_L \tag{31b}$$

where A_I is the fractional area of the oceans covered by ice. The factor A_L, being the fraction of a given latitude belt occupied by land, allows for the effect of the continents in channeling the north-south ocean currents. The average value of K_w is about 5×10^8 cm^2 s^{-1}, which agrees well with values found in the vicinity of the Gulf Stream, but is about 1 order of magnitude larger than values considered typical for the more quiescent parts of the oceans. This discrepancy was intentional and was an attempt to account partially for the neglected heat transport by vertical circulations (cf. section C3*d*).

The reflectivity and the transmissivity of the atmosphere were specified separately for a clear and a cloudy atmosphere by utilizing a two-stream approximate solution to the multiple scattering problem without zenith angle dependence. The surface albedo was defined according to the surface cover: either land, water, snow, or ice. Snow cover and ice cover were assumed to depend both on the surface temperature at the start of each month and on the average annual temperature. Thus albedo-temperature coupling was again included. An iterative method of solution was used for each month.

The original paper [*Sellers*, 1973] found at least three equilibrium climates compatible with the present value of the solar constant, corresponding to the present climate, an interglacial climate, and a glacial climate. Each was reached by starting with the specified initial conditions and running the model until in any month the surface temperature in all latitude belts over land and water did not change by more than 0.1°K from one year to the next. A recent change of this convergence criterion to 0.001°K and an integration period of about 300 model years showed all three climates to merge into one slightly colder than the present one (W. D. Sellers, per-

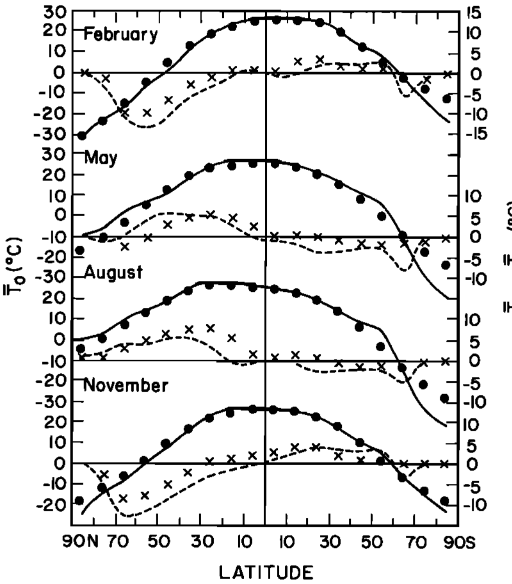

Fig. 16. Latitudinal variation of the observed mean sea level temperature (solid lines) and of the observed temperature difference between land and water (dashed lines) for February, May, August, and November. Computed values are indicated by solid dots and crosses [*Sellers*, 1973].

sonal communication, 1973). Thus it is important to be certain that a time-dependent model has reached equilibrium before conclusions as to its possible intransitivity are drawn. *Schneider and Gal-Chen* [1973, equation (9)] also discussed this point in reference to their time-dependent version of Sellers' original zonally averaged steady model.

The ability of *Sellers'* [1973] relatively simple but highly tuned model to reproduce many of the essential features of the earth-atmosphere system is indicated in Figures 16 and 17, in which computed and observed variables are compared. Figure 16 shows the latitudinal variation of the mean sea level temperature and of the temperature difference between land and water for February, May, August, and November. The observed values were derived by *Sellers* [1973], using data from other authors.

Observed and computed values of mean sea level temperatures agree well in all months everywhere except near the poles. At the south pole the computed temperatures are consistently higher than those observed, particularly in May and August during the southern hemisphere winter. This finding may simply reflect the inability of the model to account for surface inversions, since the lapse rate is specified by (30). On the other hand, extrapolation of surface temperatures from the elevation of the Antarctic Plateau to sea level is questionable at best. At the north pole the computed temperatures seem to lag slightly behind those observed.

The computed and observed temperature differences between land and water are generally of the same magnitude and sign, and in some cases (particularly at 65°S, where there is little land) Sellers asserts that 'the computed temperature differences seem more realistic than those derived from observations.' Similarly, owing to the scarcity of data on precipitation, Sellers argues that the model may be a more accurate source of precipitation data than observations.

The solid curves on Figure 17 show the much greater strength of the model-generated Hadley circulation in winter than in summer, the winter circulation being considerably stronger than is indicated by the observations of *Oort* [1971] (the dashed lines on Figure 17). Sellers suggests that this difference arises because Oort's data are 'marred somewhat by irregularities, which probably should not appear in long-term mean values. In the model, the latitude of maximum poleward transport in the Northern Hemisphere shifts from about 25°N in February to 60°N in August. At the same time, the magnitude of the transport decreases by about 60%. As might be expected, the seasonal variation is much less in the Southern Hemisphere.'

The discrepancies between Sellers' model and Oort's data probably arise from model simplifications but possibly also arise from data deficiencies or a combination of these and in particular from the fact that the averaging period may be too short or the data too few in number (see section A5, which discusses the noise problem) to allow a valid comparison between model and observations. Clearly, the signal-noise problem applies to all model observation-validation studies.

The general comments made in the previous section about highly parameterized energy balance models are also applicable here. That is, how useful can a model be for sensitivity or climate change studies if its numerous parameters have been tuned to the present climate? Although the answer to this question is not possible with the annual average energy balance models, it may be hinted at in a seasonal model, such as that of *Sellers* [1973], provided that the tuned parameters are functions at most of temperature and latitude but not of time. Since the seasons exhibit large climate changes, changes larger than some expected long-term annual mean changes, the ability of a model to reproduce the seasonal cycle with some fidelity is at least a positive indication that the model can also be used for investigations of climate change.

3. *Eddy Flux Parameterizations in Zonally Symmetric Dynamic Models*

a. Background. The energy balance models discussed in the previous sections are distinguished by the assumption

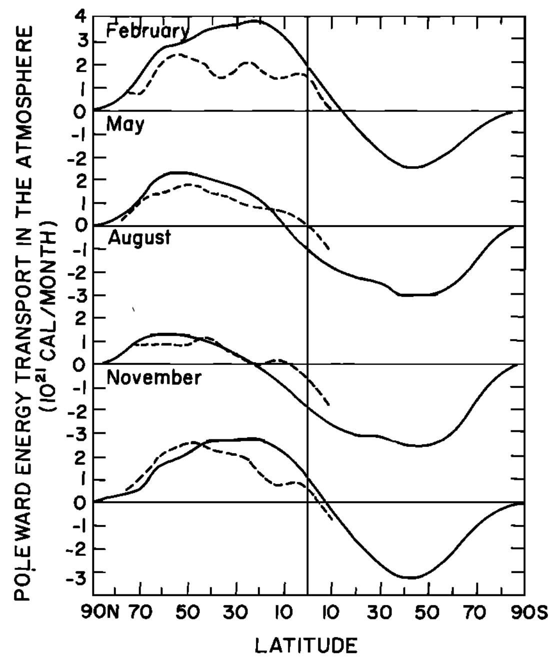

Fig. 17. Observed (dashed lines) and computed (solid lines) latitudinal variations of the total poleward energy transport in the atmosphere in February, May, August, and November. Positive values indicate a northward transport [*Sellers*, 1973].

that atmospheric temperature depends only on surface temperature. Whereas the distribution of diabatic (i.e., radiative and small-scale sensible and latent) heating may be largely controlled by local surface temperature, it is doubtful that the adiabatic redistribution of thermal energy by large-scale atmospheric motions has any straightforward surface temperature dependence. An independent calculation of atmospheric temperature should be based on the diabatic and adiabatic heating terms and requires considerable insight into relationships between various large-scale processes as provided by theoretical dynamic meteorology. Thus it is likely that parameterization of eddy flux transports will call for knowledge of many large-scale meteorological fields and not merely surface temperature. The large-scale distribution of winds, pressures, and temperatures (away from the equator) are related by geostrophic and hydrostatic balance, so that it becomes necessary to solve for the distribution of motions jointly with temperature.

We restrict our discussion in this section to longitudinally averaged dynamic models, that is, models depending on latitude and height only. For the purpose of simply viewing the more significant balances among terms of the dynamic equations consider the following 'prototype' equations:

Zonal momentum

$$\frac{\partial\langle u\rangle}{\partial t} - f\langle v\rangle + \frac{\partial\langle u'v'\rangle}{\partial y} = F \tag{32}$$

Meridional momentum
(geostrophic balance)

$$f\langle u\rangle + R\langle T\rangle \frac{\partial(\ln\langle p\rangle)}{\partial y} = 0 \tag{33}$$

Hydrostatic balance

$$\frac{\partial(\ln\langle p\rangle)}{\partial z} = -g/(R\langle T\rangle) \tag{34}$$

Thermodynamical

$$\frac{\partial\langle T\rangle}{\partial t} + \frac{\partial\langle v'T'\rangle}{\partial y} + \frac{\partial\langle w'T'\rangle}{\partial z} + \langle w\rangle\left(\frac{g}{C_p} + \frac{\partial\langle T\rangle}{\partial z}\right) = Q/C_p \tag{35}$$

Continuity

$$\frac{\partial\langle\rho\rangle\langle v\rangle}{\partial y} + \frac{\partial\langle\rho\rangle\langle w\rangle}{\partial z} = 0 \tag{36}$$

In the above equations, angle brackets denote a zonal average, and primes a deviation from the zonal average; u, v, and w are the eastward, northward, and vertical velocities, respectively; $\langle\rho\rangle = \langle p\rangle/(R\langle T\rangle)$; f is the Coriolis parameter; R is the gas constant; x and y are the distances in the eastward and northward directions, respectively; F represents frictional loss of momentum; and Q is diabatic heating. In these equations we have omitted such complications as corrections for the sphericity of the earth and other terms that are usually significantly smaller than the terms included, since our purpose is primarily to discuss eddy flux parameterizations.

Equation (32) shows that for near-steady conditions a divergence of the horizontal eddy momentum flux $\langle u'v'\rangle$ is largely balanced by the Coriolis torque term $f\langle v\rangle$. The annual mean observed distribution of $\langle u'v'\rangle$ flux is largely poleward, the maximum values being in the subtropics. This finding implies for a steady state planetary momentum balance a poleward mean meridional velocity $\langle v\rangle$ in tropical latitudes, where $\langle u'v'\rangle$ diverges, and an equatorward $\langle v\rangle$ in middle latitudes, where $\langle u'v'\rangle$ converges. From (36) this balance in turn requires mean upward vertical motion (from $(\partial\langle w\rangle/\partial z) > 0$) and hence adiabatic cooling in equatorial and high latitudes, and it requires mean downward motion and hence adiabatic heating in the subtropics, as first discussed by *Kuo* [1956].

Thus the horizontal eddy momentum fluxes serve not only to transfer zonal momentum but also to 'drive' a meridional circulation that redistributes thermal energy in the manner required for the zonal mean winds to remain in geostrophic balance with the thermal fields. Figure 18 [after *Dickinson*, 1971*a*] illustrates in terms of a simple diagnostic model the fields of zonal wind, horizontal and vertical meridional circulation, and temperature perturbation implied by a distribution of eddy momentum fluxes similar to the observed annual mean values. The indicated perturbation of the tropospheric temperature should be a significant factor in the determination of the surface balance.

In middle to high latitudes the horizontal transfer of heat by large-scale eddies, $\langle v'T'\rangle$, is significant in (35) and is thus a major factor in determining the latitudinal variation of atmospheric temperature. These generally poleward eddy heat fluxes serve to convert to eddy energy the available potential energy stored in the latitudinal variation of zonal mean atmospheric temperature. The usual theoretical interpretation of the development of these eddies is that they represent random instabilities developing on a hypothetical, zonally symmetric state (that is, symmetric about the polar axis, or having no variation in longitude). The zonally symmetric state is considered to be generated by zonally symmetric thermal forcing, such as the distribution of incoming solar energy, as would be present if there were no longitudinal variations in conditions at the earth's surface. In practice, this state is identified with longitudinal (zonal) mean variables.

It is evident from (35) that convergence of the eddy heat flux terms can be balanced by a nonzero vertical component of a mean meridional circulation, $\langle w\rangle$, as well as by diabatic heating or a change in the temperature distribution. The possible redistribution of thermal energy by the mean meridional circulation does not vanish with vertical averaging. Thus a parameterization of $\langle v'T'\rangle$ and $\langle w'T'\rangle$ alone does not provide the net latitudinal redistribution of thermal energy. Rather, it is generally necessary to consider a complete dynamic system, such as (32)–(35), to evaluate the effect of eddy fluxes.

Recent parameterizations of the eddy heat fluxes have been based explicitly on the view that these fluxes are a manifestation of the release of the zonal available potential energy by baroclinic instability. According to this approach the eddy momentum fluxes are a consequence of the stirring provided by these instabilities in the presence of the earth's vorticity and the latitudinal variation of zonal winds.

To review derivations of eddy fluxes from baroclinic wave theory, we can consider three questions (as treated separately below): (1) the development of baroclinic waves, with emphasis on the dependence of eddy fluxes on the relevant parameters; (2) the interaction of finite amplitude waves with the zonal mean wind; and (3) simple mixing approximations for the eddy transport of heat and potential vorticity, the mixing coefficients depending on the zonal mean temperature and wind structure.

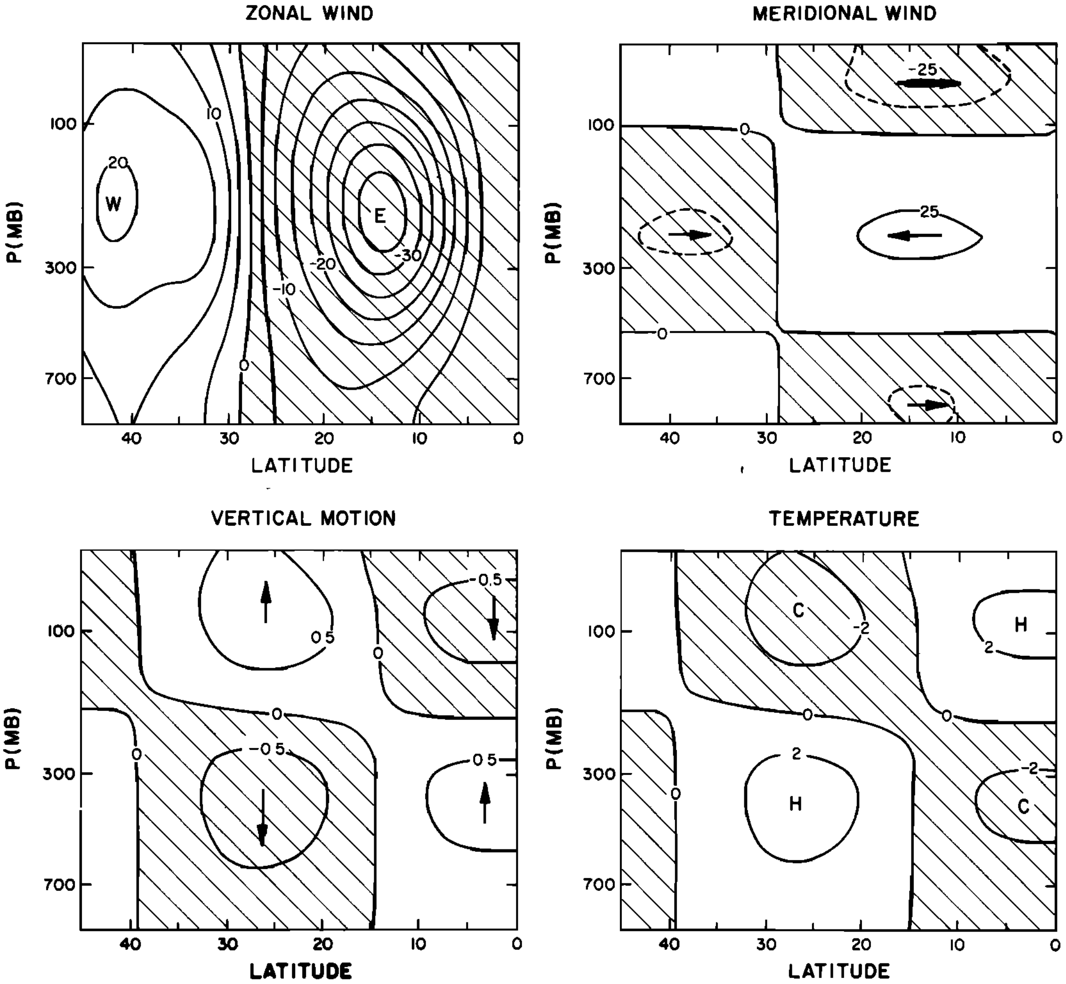

Fig. 18. Zonal mean variables calculated as response to specified distribution of eddy momentum fluxes similar to those observed. (Upper left) Zonal wind in meters per second. (Upper right) Meridional wind in centimeters per second. (Lower left) Vertical motion in millimeters per second. (Lower right) Perturbation temperature in degrees Celsius [*Dickinson*, 1971*a*].

It should be remembered that much of the observed eddy transport of heat and momentum is due to quasi-stationary planetary waves forced by inhomogeneities at the earth's surface (e.g., those discussed in section C2*d*) rather than to developing baroclinic waves (about 30% of the maximum value of the annual vertical mean heat transport is due to standing eddies [*Oort and Rasmusson*, 1971, pp. 127–128]). Consequently, very good agreement should not be expected between observed fluxes and those parameterizations derived solely from baroclinic instability processes.

b. Baroclinic wave theory. In the development of theoretical models for latitudinal eddy transports of heat it is generally assumed that the thermal and wind fields are nearly in hydrostatic and geostrophic balance, so that the eddy horizontal velocities and temperature can be approximately represented by a stream function ψ. For example,

$$\begin{aligned} u' &= -\partial\psi/\partial y \\ v' &= \partial\psi/\partial x \\ T' &= (f\langle T\rangle/g)\,\partial\psi/\partial z \end{aligned} \tag{37}$$

Furthermore, it is often assumed that the spherical geometry can be approximated with a middle latitude 'tangent plane' where the Coriolis parameter f is held constant except where it is differentiated with respect to y. If it is also assumed that $u' \ll \langle u\rangle$, $T' \ll \langle T\rangle$, and $p' \ll \langle p\rangle$, so that primed quantities represent a small perturbation on a mean zonal velocity that is in geostrophic balance with the zonal mean temperature field, then the governing equations for the motion and temperature perturbations reduce to a single equation for the conservation of potential vorticity q, which for inviscid adiabatic flow (no friction or heating) is written as

$$\begin{aligned} \frac{Dq'}{Dt} + v'\frac{\partial\langle q\rangle}{\partial y} &= 0 \\ q' &= \left[\frac{1}{\langle\rho\rangle}\frac{\partial\langle\rho\rangle}{\partial z}\frac{f^2}{N^2}\frac{\partial}{\partial z} + \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2}\right]\psi \\ \frac{\partial\langle q\rangle}{\partial y} &= \beta - \frac{1}{\langle\rho\rangle}\frac{\partial}{\partial z}\langle\rho\rangle\frac{f^2}{N^2}\frac{\partial\langle u\rangle}{\partial z} - \frac{\partial^2\langle u\rangle}{\partial y^2} \end{aligned} \tag{38}$$

and where $\beta = df/dy$ and N is the buoyancy frequency, derived from $\langle T(z)\rangle$. The 'substantial' derivative D/Dt is approximated by

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \langle u\rangle\frac{\partial}{\partial x} \tag{39}$$

Equation (38), with the boundary condition that vertical motion vanishes at the lower boundary, serves to describe the evolution of small-amplitude (primed) disturbances on the zonal flow $\langle u(y, z)\rangle$. The zonal flow is said to be unstable to these disturbances if such a disturbance taken as an initial condition grows with time. The instabilities of greatest interest are those with exponential growth, so that it is customary to examine the growth of individual Fourier components, i.e.,

$$\psi = \Psi(y, z) \exp [i\alpha(x - ct)] \tag{40}$$

which, together with (39) substituted into (38), reduces it to a boundary value partial differential equation in y and z:

$$\left[\frac{1}{\langle\rho\rangle}\frac{\partial}{\partial z}\frac{f^2}{N^2}\langle\rho\rangle\frac{\partial}{\partial z} + \frac{\partial^2}{\partial y^2} + \frac{\partial\langle q\rangle/\partial y}{\langle u\rangle - c} - \alpha^2\right]\Psi = 0 \tag{41}$$

Theoretical studies of (41) are usually simplified further with the assumption that $\langle u(y, z)\rangle$ is a function only of z or only of y. The resulting equations are referred to as the 'baroclinic' or 'barotropic' stability problems, respectively. North-south zonal temperature gradients require variation in the vertical direction of the zonal wind, as is shown by (33) and (34). Instability to small disturbances of a zonal wind with variation only in the vertical (baroclinic instability) direction converts to eddy energy the energy that is stored in the latitudinal variation of zonal temperature. This energy is released by eddy flux of heat $\langle v'T'\rangle$ from higher to lower mean temperatures, that is, in the poleward direction in middle latitudes. Instability of the zonal wind to horizontal shears, on the other hand, converts kinetic energy of the zonal wind to eddy energy through an eddy flux of momentum $\langle u'v'\rangle$ from regions of greater to regions of lesser zonal momentum. When both vertical and horizontal shears are present, the mechanism of instability is evaluated according to the direction of energy conversions. For example, if a disturbance transports heat to colder zonal temperatures but momentum to regions of a higher zonal momentum, it is said to be baroclinically unstable but barotropically stable. If these definitions are generalized to the direction of energy flow for actual atmospheric finite amplitude zonal asymmetries, then climatological data show the tropospheric zonal flow as given by observations to be that of baroclinic instability and barotropic stability (i.e., depending upon the direction of energy flow) [e.g., *Newell et al.*, 1969].

Equation (41) is solved as an eigenvalue problem for the complex phase speed c. Solutions with the imaginary part of c greater than zero represent unstable modes. The other most significant parameters of the problem are the longitudinal wave number α and the strength of wind shear. In the simplest model of baroclinic instability the atmosphere is approximated by two layers in the vertical direction with differing zonal velocities, the difference of which is characterized by a shear parameter. Equation (41) then reduces to two linear equations that are solved algebraically for the complex phase speed c. For sufficiently long or short wavelengths or for sufficiently weak shear, both roots are real, and so there is no unstable mode. For large enough vertical shear and an intermediate range of wavelengths there are complex conjugate roots and hence one growing mode. The region of unstable modes and their growth rates depending on these parameters are shown in Figure 19 [after *Phillips*, 1954]. The baroclinic waves of fastest growth have wavelengths between 5000 and 8000 km according to Figure 19. These waves are associated with the mid-latitude cyclonic storms, which are also typically of this scale. The perturbation temperature and the northward velocity are partly in phase, giving a mean poleward eddy heat transport, which provided that wave amplitudes are assumed comparable to those of the observed eddies, is similar in magnitude to the observed poleward heat transports. Typical shears are 2–4 m s^{-1} km^{-1}.

For greater validity it is desirable to examine baroclinic waves with much more vertical resolution than is provided by two layers. *Eady* [1949] assumed a stratified but incompressible model of the atmosphere with a zonal wind that was linear in altitude up to some top lid and was without the β term of (41). Under these conditions he showed that a simple complex trigonometric solution could be derived that represented a growing mode for wavelengths longer than a critical wavelength. For the more realistic case of an unbounded stratified atmosphere on a β plane *Charney* [1947] and *Kuo* [1952] established the existence of a neutral solution, as given by the dotted curve shown in Figure 19, and a growing mode with growth rates similar to those shown in Figure 19 for shorter wavelengths. However, they found no short-wavelength limit to instability but rather small growth rates outside the region of two-layer instability. *Green* [1960] showed numerically that there are also slowly growing modes on the long-wavelength side of the dotted line, so that the β plane model with continuous shear is unstable for all shears and wavelengths excluding certain curves in the parameter space on which neutral solutions lie. *Miles* [1964] established analytically that this model is always unstable for sufficiently short wavelengths or small wind speeds. The presence or absence of slowly growing modes is generally regarded as being of little physical significance if there are also much more rapidly growing modes present. This improved mathematical understanding of the β plane baroclinic instabilities would not be worth mentioning in the present context except for the fact that it has allowed profound physical insight into the nature of the problem, suggesting, for example, an explanation for the direction of eddy momentum transport provided by growing baroclinic waves in the presence of horizontal shear.

Charney and Stern [1962] show that with absent (i.e., no bottom boundary) or constant temperature horizontal boundaries, instability is only possible if the following integral constraint is satisfied:

$$\iint_{\text{domain}} \langle\rho\rangle \, dz \, dy\left[\frac{|\psi|^2 \, \partial\langle q\rangle/\partial y}{|\langle u\rangle - c|^2}\right] = 0 \tag{42}$$

which is possible only if the gradient of potential vorticity changes sign over the domain. *Bretherton* [1966] interpreted

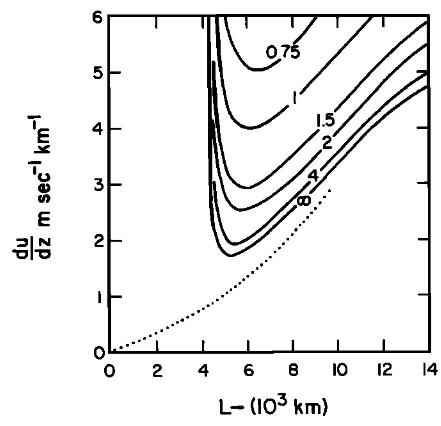

Fig. 19. Time in days required for an unstable wave in the two-level model to double its amplitude, given by *Phillips* [1954] as a function of the vertical wind shear in the basic current and the wavelength. The dotted line represents the curve $\sigma = 0$ for the perturbation analyses of *Charney* [1947] and *Kuo* [1952].

this constraint as being equivalent to a requirement that eddy transport of potential vorticity must be of different signs in different regions for instability to be possible. He showed that the bottom boundary condition with a mean temperature gradient is equivalent to an infinitesimal sheet of potential vorticity gradient at that level, negative in value where the temperature decreases poleward. The observed values of $\langle q_y \rangle$ are generally positive everywhere in the troposphere (except possibly near the poles) because of the dominance of β (cf. definition (38)). Hence it is the equivalent sheet of potential vorticity gradient at the earth's surface that is required for baroclinic instability to occur in the troposphere. Bretherton pointed out that at critical levels (where $u \simeq c$) the potential vorticity flux becomes nearly independent of growth rates, whereas elsewhere in the domain it is proportional to growth rates. Thus if an otherwise neutral wave reaches a critical level, it must grow or decay to allow a return flux of potential vorticity elsewhere, e.g., at the lower boundary. Hence the neutral waves of the two-layer model in the presence of continuous shear and a critical level break up into two modes with complex conjugate phase speeds and hence one unstable mode.

McIntyre [1970] has carried this line of analysis further, assuming the Eady model to a first approximation, perturbing it with a horizontally varying jet, and showing that eddy momentum fluxes $\langle u'v' \rangle$ tend to strengthen the jet. *Stone* [1969] has developed another type of perturbation expansion, obtaining similar results by considering a two-layer β plane model with horizontal shear.

c. Finite amplitude wave-zonal flow interaction. As a baroclinic wave grows in the presence of a mean zonal wind, it provides an eddy transport of heat and momentum proportional to the squared amplitude of the wave. These transports are in turn expected to change the zonal flow. These changes will feed back on the structure of the wave and generally limit its growth. To study this feedback between baroclinic waves and zonal flow, it is necessary to consider not only a perturbation potential vorticity equation, e.g., (38), but also mean flow equations, e.g., (32)–(36), which as written can be reduced to a single equation for the zonal mean potential vorticity. When a single longitudinal Fourier component is considered, the mean and perturbation equations depend only on latitude and time. This coupled system has been solved by *Simons* [1972] with realistic zonal winds. Figure 20, taken from that study, shows for the small-amplitude modes in his study the variation with longitudinal wave number of the baroclinic conversion from zonal available potential energy by eddy heat fluxes and barotropic conversion to zonal kinetic energy by eddy momentum fluxes. Simons finds that the barotropic energy exchange (at least under the conditions of his study) plays a major role in limiting the growth of baroclinic waves.

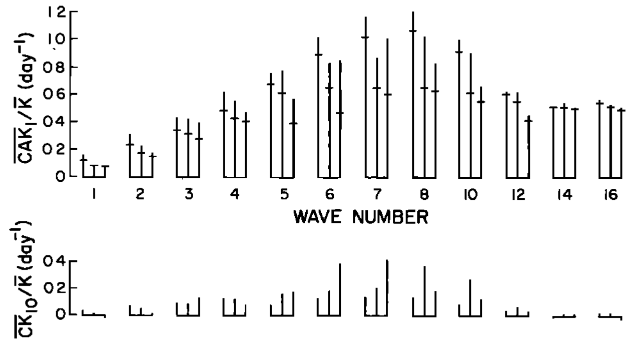

Fig. 20. Energy conversions of the three most unstable modes for each wave number. (Top) Conversion of potential to kinetic energy. The net rate of change of wave kinetic energy is represented by the parts of the vertical bars below the horizontal marks [after *Simons*, 1972]. (Bottom) Release of kinetic energy from the wave to the zonal flow.

To develop a more analytic approach to the problem of a baroclinic wave interacting with a zonal flow, *Pedlosky* [1970, 1971] has assumed a two-layer model with only vertical shear and a wave near enough to marginal stability that its growth time is large in comparison with its oscillation period. Eddy heat fluxes driving the zonal flow then depend only on the wave amplitude and a slow rate of change in wave amplitude, whereas an equation for wave amplitude is derived that depends essentially on the change of instability growth rates by the changed zonal flow. These equations simplify in his perturbation expansion to a single first-order ordinary differential equation in time for the wave amplitudes. He is able to find steady, finite amplitude wave solutions as well as waves whose amplitudes oscillate in a limit cycle. *Drazin* [1972] has carried out a similar analysis for the Eady continuous shear problem.

Another approach has been pursued by *Kurihara* [1970, 1973], who uses a two-layer primitive equations model (that is, one assuming no approximate geostrophic balance) and attempts to derive directly time-dependent equations for zonal mean second-moment eddy quantities (i.e., fluxes and variances) in terms of the zonal mean temperature. The eddies are not assumed to be single Fourier waves, but Kurihara replaces Laplacian operators with $-l^{-2}$, where l is a length scale, an approximation of unclear significance unless the wave is given by a single spherical harmonic component. Eddy momentum flux is assumed in the lower layer to be absent and in the upper layer to be effectively carried downward by Coriolis torques to exactly balance surface drag on the zonal wind. Thus the eddy momentum flux depends only on the surface zonal wind. Eddy heat flux comes out of several time-dependent equations solved jointly for eddy and mean flow amplitudes. For a steady state it must satisfy the constraint that generation of eddy available potential energy by poleward eddy heat fluxes balances the loss of kinetic energy by eddies to the zonal flow by upgradient momentum transport. Solutions for the model appear to reproduce general features of the zonal atmospheric flow. However, the advantages of this approach over the use of a single Fourier wave are not clear, and insight into details of the model dynamics is more difficult to achieve.

The wave-zonal flow interaction studies discussed up to now have involved waves with amplitudes that are growing or at least changing in time. The generation of time-averaged statistics then requires an integration over time. An alternative approach would be to assume that (seasonal variation being excluded) the eddy statistics should be stationary in time. Then the problem is to relate the time-averaged statistics to the zonal state. In one such study *Dickinson* [1969] showed that statistically stationary eddies must transport potential vorticity in order to change the zonal flow and that this can occur only in the presence of wave critical levels (i.e., where $\langle u \rangle = c$) or of dissipation. This conclusion may no longer be valid if zonal winds oscillate on the same time scale as the waves.

d. Eddy mixing by baroclinic waves. The previous section dealt with the task of parameterizing eddy transports in terms of statistical averages taken over the complete life history of baroclinic waves. It is somewhat simpler to neglect the decay phase of these waves and to assume that the eddy fluxes are determined primarily by the structure of small-amplitude growing waves as given by linear stability theory. This approach is now reviewed.

Early studies [e.g., *Williams and Davies*, 1965; *Dolzhanskiy*, 1969] have heuristically argued that since baroclinic waves are driven by the meridional temperature gradient, their eddy transports are most simply parameterized as being proportional to this gradient, i.e.,

$$\langle v'T'\rangle = -K_T\, \partial\langle T\rangle/\partial y \tag{43}$$

$$\langle u'v'\rangle = -K_M\, \partial\langle T\rangle/\partial y \tag{44}$$

where K_T and K_M are coefficients to be selected from observation. More recently, various authors [e.g., *Green*, 1970; *Saltzman and Vernekar*, 1971; *Stone*, 1972*a*, 1973] have attempted to derive parameterizations from mechanistic models of baroclinic waves (see section D3*b*). Although the detailed arguments differ, the forms of the parameterizations of eddy heat transport suggested by the above studies are rather similar.

The role of the baroclinic wave model is primarily to relate the amplitude of eddy temperature to eddy northward velocity (equation (43)). If the wave evolves in time as exp (σt), where σ is the growth rate, and if the rate of temperature growth is essentially balanced by eddy advection of the mean temperature field, then the perturbation temperature equation can be reduced to

$$T' = (1-\epsilon)\frac{\partial\langle T\rangle}{\partial y}v'/\sigma \tag{45}$$

where $\epsilon < 1$ allows for some cancellation of the temperature advection by vertical motion adiabatic cooling. More generally, there will be besides (45) another part of T' out of phase with v' and therefore not available (in the zonal mean) for transport.

From (45)

$$\langle v'T'\rangle = (1-\epsilon)\langle v'^2\rangle/\sigma\ \partial\langle T\rangle/\partial y \tag{46}$$

From the baroclinic instability theory of Eady the most unstable waves have growth rates of the form

$$\sigma = O(1)fRi^{-1/2} \tag{47}$$

where Ri is the Richardson number,

$$Ri = N^2/(\partial\langle u\rangle/\partial z)^2 \tag{48}$$

and $O(1)$ denotes a quantity of the order of unity. Hence (46) may be written as (43) with

$$K_T = O(1)\langle v'^2\rangle\frac{N/f}{\partial\langle u\rangle/\partial z} \tag{49}$$

where $(1-\epsilon)$ is lumped into the $O(1)$ factor. Thus an expression for K_T is obtained that is proportional to the variance of north-south velocities and inversely proportional to the vertical wind shear (or horizontal temperature gradient). In order to achieve a K_T with wind shear in the numerator *Stone* [1972*a*, equations 2.16–2.21] argues that the energy conversion processes should give eddy energies comparable in magnitude to that of the zonal flow kinetic energy, and by referring to observations he infers that in a vertically averaged sense,

$$\langle v'^2\rangle \simeq (H\ \partial\langle u\rangle/\partial z)^2 \tag{50}$$

where $H = R\langle T\rangle/g$ is atmospheric scale height. Hence he replaces (49) with

$$K_T = O(1)(N/f)H^2\ \partial\langle u\rangle/\partial z \tag{51}$$

To express (51) in terms of thermal gradients, he uses a thermal wind expression, essentially

$$\frac{\partial\langle u\rangle}{\partial z} = -\frac{1}{fH}\frac{\partial(R\langle T\rangle)}{\partial y} \tag{52}$$

so that

$$K_T = O(1)NH/f^2\frac{\partial}{\partial y}(R\langle T\rangle) \tag{53}$$

Taking as typical values $N \simeq 10^{-2}$ s^{-1}, $f \simeq 10^{-4}$ s^{-1}, $H \simeq 10^6$ cm, $\partial(R\langle T\rangle)/\partial y \simeq 2\cdot 10^{-2}$ cm s^{-2}, we have

$$K_T = O(1)\cdot 2\cdot 10^{10} \tag{54}$$

This derivation is essentially the same as Stone's except that by reference to the Eady model Stone is able to establish values for the $O(1)$ factors. In particular, Stone finds $O(1) = 0.722$ as the appropriate factor in (49), (51), and (53). *Saltzman* [1968*b*] suggested an expression similar to (49) but in which Stone's assumption (50) is replaced by the nearly equivalent assumption $\langle v'^2\rangle \simeq \sigma^2$ and in which (47) is inferred as an approximation from studies of *Charney* [1947] and *Kuo* [1952]. In their latter study *Saltzman and Vernekar* [1971] assumed instead that in (49), $\langle v'^2\rangle \simeq \sigma$, thus removing the linear dependence of K on the meridional temperature gradient in (53). The requirement that on a global mean, eddies must provide sufficient upward heat transport to balance upper-level radiative cooling is used to establish the value of a proportionality constant and to give an expression for K_T [cf. *Saltzman and Vernekar*, 1971, equation (77)] that appears to depend on the global mean Richardson number (equation (48)) and hence inversely on the square of the horizontal temperature gradient. This assumption seems to require that the vertical eddy heat transport adjust to whatever the upper-level cooling is. It is not clear, however, whether the atmosphere actually behaves in this manner or whether conversely, the upper-level cooling adjusts through a change in temperature until it balances whatever upward heat fluxes occur.

Stone [1972*b*] shows that under the conditions of the Eady model for baroclinic waves, mean vertical motion vanishes; i.e., $\langle w\rangle = 0$. Consequently, he is able to use his heat flux parameterization alone for a consistent, albeit somewhat incomplete, climate model (e.g., no ice feedback or moisture is included). It is always interesting to see what the consequences are of such simple models with only one mechanism for latitudinal and vertical thermal coupling. *Stone* [1972*a*, 1973] averages in latitude and vertical direction to obtain simple equations for predicting global mean climatic conditions in terms of various physical parameters. *Stone* [1972*a*] shows how vertical eddy heat transport can stabilize what would be, with only radiative heating, a convectively unstable temperature profile. For terrestrial conditions these fluxes alone maintain a lapse rate 2°K less than the adiabatic rate (whereas the U.S. Standard Atmosphere lapse rate is 3.5°K less than the adiabatic rate).

Stone [1972*a*] solves the global mean climate model defined by his averaged equations and studies [*Stone*, 1973] the varia-

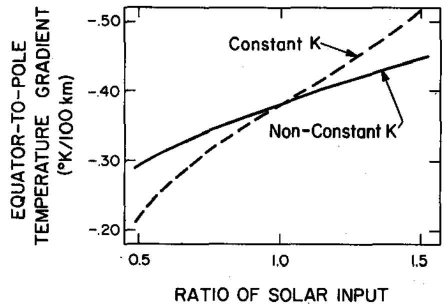

Fig. 21. Equator-to-pole temperature gradient versus input of solar energy as derived by *Stone* [1973] for the case of a constant eddy diffusion coefficient K and for Stone's parameterization for K that depends upon the temperature gradient.

tion of the solutions with variation of external parameters. Increased solar input in his climate model and hence increased radiative differential heating generate an increase in the magnitude of the latitudinal temperature gradient as shown in Figure 21 [after *Stone,* 1973]. The dotted line represents a constant value of the eddy diffusion coefficient K, and the solid line represents results from using the nonconstant K of *Stone* [1973], which is a function of the latitudinal temperature gradient. Thus in the absence of other feedbacks Stone shows that the assumption of a constant eddy diffusion coefficient leads in his model to an overestimation by a factor of 2 of the variation of latitudinal temperature gradient with solar constant changes.

Stone concludes that *Sellers'* [1969] (e.g., our section D2*a*) 'estimate of how large a decrease in solar constant is necessary to initiate an ice age is low because of his assumption of a constant K.' This supposition of Stone has recently been confirmed numerically by *Gal-Chen and Schneider* [1974], who use Stone's nonlinear prescription for K in their time-dependent version of *Sellers'* [1969] model. *Sellers* [1973] himself has used a parameterization equivalent to Stone's but has not discussed comparative effects. The static stability of Stone's model changes very little even with large changes in the solar constant. This finding is attributed (in the case of an increase in solar input) to a near balance between the increase of radiative destabilization with increased surface temperature and the increase of vertical eddy flux stabilization with increased poleward eddy heat transport (consequent to the increase of mean temperature gradient shown on Figure 21).

It should be remembered that in order to isolate the possible role of eddy heat flux Stone's model has deliberately omitted other equally important climatic feedback mechanisms, in particular, condensational heating and ice albedo-temperature coupling (cf. section B2*c*). Thus the conclusions of his climate model should be interpreted primarily as illustrating the importance of the eddy heat fluxes and should not be regarded as approximations to reality. For example, climate models with ice albedo-temperature coupling [e.g., *Sellers,* 1969, 1973] predict that a change in solar radiation will produce the largest temperature changes in polar latitudes and that consequently, the change of latitudinal temperature gradient will be the opposite of that calculated by *Stone* [1973]. The GFDL GCM has recently been used by S. Manabe (private communication, 1974) and has also been described by *Smagorinsky* [1974] to test the response of his model to uniform increases and decreases in solar input. The GCM's are three-dimensional dynamic models that compute eddy flux transports in detail, and since this version of the GFDL model computes the temperature of the earth's surface and accounts for (among other things) snow albedo feedback and temperature-lapse rate feedback (also see our section B2), it is appropriate to compare Manabe's result in Figure 22 (also reprinted as Figure 18.35 of *Smagorinsky* [1974]) with the results of the highly parameterized models of Sellers and Stone. From Figure 22 it is clear that an increase of 2% in solar input has caused an increase in surface temperature everywhere, the largest increase being at the poles. Thus the GCM simulation supports the conclusion that snow albedo feedback and temperature-lapse rate coupling dominate eddy flux transport processes and result in a decrease in the equator-to-pole temperature gradient from an increase in the solar constant (as also found by Sellers and in contrast with Figure 21, adapted from Stone). However, this decrease in gradient is by no means uniform with latitude. S. Manabe (private communication, 1974) also found that the converse was true for decreases in the solar constant, except that the temperature contrast was even greater (as is also true for the semiempirical theories of *Budyko* [1969] or *Sellers* [1969]).

Stone offers the results from his model as an explanation of the near constancy of vertical lapse rates. Again, we may refer to Manabe's results shown in Figure 22, which suggest that the hemispheric average lapse rate is decreased by a surface temperature increase, especially in the tropics, where moist convection is important. This conclusion was also drawn by experiments with the NCAR GCM, and its implications for cloudiness feedback (not included in Manabe's runs) were discussed in sections A5 and B2*g* of this paper. It seems likely that with the addition of ice albedo-temperature coupling to Stone's eddy heat flux model the sign of the feedback changes, and we will be forced to conclude that vertical eddy heat fluxes act to enhance rather than to dampen changes in lapse rate originating from changes in radiation. That is, if solar input is increased in a model including radiation and eddy heat

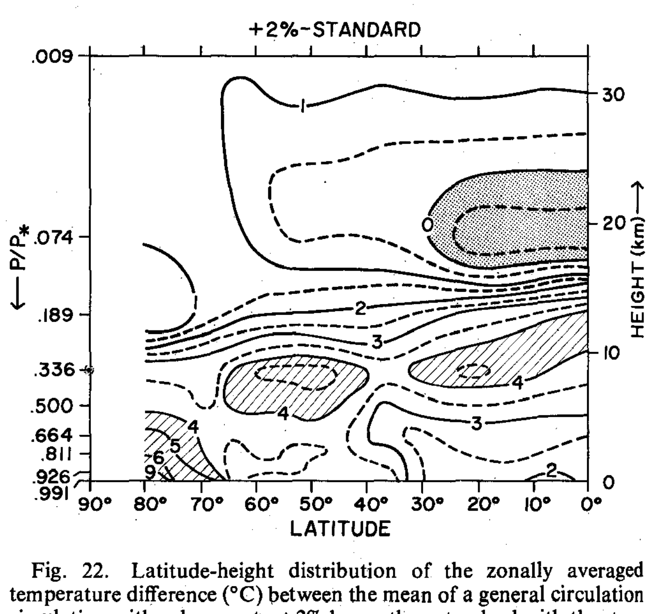

Fig. 22. Latitude-height distribution of the zonally averaged temperature difference (°C) between the mean of a general circulation simulation with solar constant 2% larger than standard with the standard. This is a fully three-dimensional model with idealized geography and also includes coupling between surface albedo and snow cover as generated by the model. Note the decrease in equator-to-pole surface temperature gradient (as contrasted with Figure 21) and the general decrease in lapse rate (S. Manabe, private communication, 1974) (also to appear as Figure 18.35 of *Smagorinsky* [1974]).

fluxes, this radiation change will in the absence of moist convection and ice feedback tend to increase both the lapse rate γ and the equator-to-pole temperature gradient $\partial T/\partial y$. The increased $\partial T/\partial y$ leads to increased horizontal eddy flux transport $\langle v'T'\rangle$, which according to baroclinic theory tends to enhance vertical transport of eddy energy $\langle w'T'\rangle$. These vertical eddy fluxes may then oppose the radiative destabilization of γ and (according to the results of Stone) nearly balance it, so that static stability is effectively unchanged by changes in solar input. However, since this balance (which maintains constant γ in Stone's model) depends upon the assumption that increased solar input yields increased $\partial T/\partial y$ (as is true in Figure 21) and consequently increased $\langle w'T'\rangle$, it would seem that constant γ would no longer be possible with snow albedo coupling (which leads to decreased $\partial T/\partial y$ with increased solar constant). If increased solar input (to a model now containing ice feedback) yields decreased $\partial T/\partial y$ and thus decreased $\langle w'T'\rangle$, then the radiative destabilization of γ will be enhanced and will not be opposed by the decrease in $\langle w'T'\rangle$!

The GFDL GCM result in Figure 22, however, shows an increase in average stability (i.e., a less steep lapse rate) resulting from an increase in solar input, whereas the arguments summarized above, based on radiative heating and vertical eddy fluxes alone, will lead to destabilization (i.e., a steeper lapse rate). Thus it seems that neither of these latter processes has a dominating influence on the lapse rate in the models that include them both, particularly in the tropics. Rather, the lapse rate in the GFDL GCM appears to be controlled primarily by moist convection—at least in the tropics. A similar conclusion is made regarding NCAR GCM simulations described in section B2*g*. However, Manabe's result (our Figure 22) is less conclusive in mid-latitudes, precisely the zone in which the baroclinic eddy influences described by Stone would be most important. This discussion illustrates the need to check inferences from simple models against more elaborate numerical simulations, which in turn can more easily be interpreted by mechanistic models such as the model of *Stone* [1973].

The simpler baroclinic models just discussed neglect variation of parameters with latitude. Inferred parameterizations of eddy heat flux have then been applied in a somewhat ad hoc fashion either in terms of local meridional temperature gradients [e.g., *Saltzman and Vernekar,* 1971] or in terms of global averages [e.g., *Stone,* 1973].

It seems likely that since even simple, relatively unrealistic models of baroclinically unstable eddies provide parameterizations of eddy heat fluxes of use to climate theory, more realistic baroclinic instability theories should be useful in extending and improving this parameterization. With this in mind *Stone* [1974*a*] discusses modes from a two-layer β plane model obtained by assuming a slowly varying horizontal shear [*Stone,* 1969]. By examining the latitudinal variation of eddy heat fluxes from several modes he infers that these heat fluxes will depend approximately on the local degree of baroclinic instability. He finds that poleward heat fluxes are generally maximum at the latitudes where the rate of poleward temperature decrease is maximum and that they change sign where locally, the simple latitude-independent stability model predicts marginal stability. The two-layer β plane model (cf. Figure 19) predicts the cessation of baroclinic instability at a finite positive value of vertical shear—in contrast to the Eady model (with $\beta = 0$), which is unstable except for zero shear. This stabilizing effect of β is approximately retained in more continuous models with continuous vertical shear, as is discussed in section D3*b*. Therefore the parameterization (53) becomes doubtful for sufficiently small $(-\partial\langle T\rangle/\partial y)$. Stone interprets these two-layer β plane criteria for baroclinic instability as a condition on the minimum slope of isentropic surfaces that permits instability, which is zero in the Eady model. He accordingly modifies his parameterization for eddy fluxes so that they reach zero for the β plane conditions of marginal stability. To do this, he multiplies K_T in (53), here denoted $(K_T)_{\text{Eady}}$, by a factor

$$\frac{(K_T)_{\beta\ \text{plane}}}{(K_T)_{\text{Eady}}} = 1 - \frac{1.42(z/z_T)[1-(z/z_T)]}{1-2.5(z/z_T)[1-(z/z_T)]}\frac{H}{R}\frac{\partial\langle\theta\rangle/\partial z}{|\partial\langle\theta\rangle/\partial y|} \tag{55}$$

where θ is the potential temperature, z_T is the altitude of the tropopause, H is the atmospheric scale height, R is the radius of the earth, and $\partial\langle\theta\rangle/\partial y$ and $\partial\langle\theta\rangle/\partial z$ are mass-weighted vertical averages of these quantities; and where it is recalled that $(K_T)_{\text{Eady}}$ was intended to apply to vertically integrated heat transports. The factor (55) is a maximum at the surface, decreases to a minimum at 400 mbar, and then increases again to the tropopause. With (55) Stone obtains approximate agreement with the vertical, latitudinal, and seasonal variation of eddy heat fluxes observed by *Oort and Rasmusson* [1971]. The maximum of predicted fluxes is always at the latitude of maximum temperature gradient. This latitude is 15° too far south in January in relation to the observed fluxes. The agreement of Stone's parameterization with observations is remarkable when it is recalled that a significant fraction of the observed fluxes is a consequence of stationary eddies that are of an origin and structure different from those of the growing baroclinic waves. (The heat transport from stationary planetary waves, which carry energy upward, is poleward and independent of the north-south thermal gradient [e.g., *Matsuno,* 1970].) However, Stone has argued that this agreement can be explained in part because of the limitations of the size of the large-scale eddies, regardless of whether they are forced stationary waves or transient baroclinic waves.

Some support for this viewpoint is provided by a comment from S. Manabe (private communication, 1974): 'Manabe and Terpstra (1974) discussed the relative importance of stationary and transient eddies in transporting heat in a model atmosphere. According to their results, the contribution of stationary eddies is very important in the model with mountains, whereas that of transient eddies becomes important in the mountainless model. However, the total eddy heat transport (i.e., the sum of these two contributions) is affected little by the effects of mountains. This result indicates that transient eddies can replace stationary eddies in transporting heat.' A particularly interesting consequence of Stone's eddy flux parameterization is the prediction of equatorward eddy heat fluxes in polar latitudes.

In view of the major role of the surface mean temperature gradient in the development of baroclinic instability (as discussed in section D3*b*) it seems that the $\partial\langle T\rangle/\partial y$ entering into the transport parameterization might be evaluated more appropriately at the surface than as a mass-weighted average, as Stone suggests.

It is again recalled that a description of eddy heat flux must be supplemented by a description of the redistribution of thermal energy by the mean meridional circulation. This description in turn requires a knowledge of eddy momentum transport. As was mentioned in section D3*c*, the net effect of eddy heat and momentum transports on zonal structure can be described (within the approximations of quasi-geostrophic theory) by zonal mean eddy transport of potential vorticity [e.g., *Green*, 1970, section 7]. Potential vorticity has the useful property that it is conserved under inviscid adiabatic motion and thus is a good candidate as a parameter to be mixed down its mean gradient by eddies. Thus the prescription suggested by *Green* [1970] is essentially to parameterize eddy heat transport in terms of an eddy mixing expression, as inferred from a baroclinic wave model, and then to use the resulting mixing coefficient to parameterize potential vorticity transport in terms of eddy mixing. The transport of potential vorticity is then used with the eddy heat transport to drive the mean meridional circulation and eddy momentum flux. Thus Green derives the net adiabatic heating and rate of zonal momentum addition by eddy fluxes and mean meridional circulation.

Green derives an expression for K_T similar to (53), using energy arguments based on the idea that $\langle v'^2 \rangle$ will be proportional to the potential energy available to a parcel, which in turn is proportional to $(\partial \langle T \rangle / \partial y)^2$. These arguments bolster confidence in the capability of (50) to reflect climate change beyond Stone's reference to observations. Furthermore, since Green's derivation of K_T does not depend on a particular small-amplitude baroclinic wave model, it lends greater support to the approximate correctness of (53). The horizontal mixing coefficient should be equal to the correlation between northward velocities and northward displacements. This correlation leads directly to an expression for the mixing coefficient of any conserved quantity as

$$K = \frac{\pi}{2} \frac{\sigma \langle v'^2 \rangle}{\alpha^2 \, |\langle u \rangle - c|^2} \tag{56}$$

where c is the wave speed. Since unstable waves generally have a phase speed near the value of $\langle u \rangle$ at the surface, (56) provides another derivation besides Stone's that predicts a vertical variation of K with a maximum near the surface. However, in contrast with Stone's expression (55), (56) will not generally have a secondary maximum near the tropopause.

The 'critical' level at which $\langle u \rangle \simeq c$ is referred to by Green as the 'steering level.' The mixing becomes concentrated entirely at that level as the baroclinic growth rate σ approaches zero. This limiting case of mixing as applied to statistically stationary waves has been analyzed in detail by *Dickinson* [1969].

Further progress in developing an eddy mixing parameterization in terms of potential vorticity transport has been made by *Wiin-Nielsen and Sela* [1971], who calculated the latitudinal transport of potential vorticity from data covering a full year. Figure 23 shows the winter averages that they thus obtained. The transport above the 800-mbar level is everywhere equatorward. Mean potential vorticity increases poleward everywhere in the troposphere because of the dominance of the Coriolis term. Consequently, the potential vorticity mixing above 800 mbar is everywhere down the mean gradient. Wiin-Nielsen and Sela interpret the northward transport below the 800-mbar level as resulting from planetary boundary effects. It may be further speculated that this transport arises from a smearing-out of the effective thin (delta function) layer of potential vorticity gradient at the surface due to the lower boundary condition (see the discussion in section D3*b*). *Sela and Wiin-Nielsen* [1971] have shown that the empirically inferred annual mean parameterizations for potential vorticity and temperature mixing coefficients can be used with a two-layer model of the zonal mean equations to derive the gross features of the seasonal variation of zonal winds and temperatures away from the tropics.

It is not necessary to parameterize the transport of both heat and potential vorticity if a direct parameterization of eddy momentum transport can be derived. This parameterization has been attempted by *Saltzman and Vernekar* [1968], who argue that

$$\langle u'v' \rangle = t_c \langle v'^2 \rangle \cos\theta \; d\mu/d\theta \tag{57}$$

where μ is the local angular phase speed of an eddy of given wave number and t_c is a characteristic time for wave tilting. Strictly speaking, (57) should be averaged over the spectrum of wave numbers, but Saltzman and Vernekar apply it with a typical value, e.g., longitudinal wave number 6. The parameter t_c is then chosen so that the eddy kinetic energy will convert to zonal kinetic energy at the rate that the zonal kinetic energy is being dissipated.

Saltzman and Vernekar [1971, 1972] have used this parameterization in a rather elaborate two-layer global climate model for surface temperature and atmospheric structure. Up to the present time none of the eddy flux climate models that we have referred to that include detailed treatments of the zonal mean dynamics have included the feedback between ice albedo and temperature. Consequently, we have not discussed any of their other details. It should be noted, however, that Saltzman and Vernekar have included self-consistently the release of latent heat in the tropical Hadley circulation. According to *Dickinson* [1971*a*, *b*] this thermal source is essential for a realistic derivation of zonal winds and temperatures.

E. CONCLUDING REMARKS

1. *Other Topics Related to Climate Modeling*

At least two further topics merit consideration in surveying the field of climate modeling, i.e., modeling of the climates of (1) the upper atmosphere and (2) planetary atmospheres. Enlarging the scope of climate modeling to encompass such subjects provides the perspective and long-range view necessary to insure that fundamental questions are given adequate attention.

a. Upper atmosphere. Above the tropopause it generally

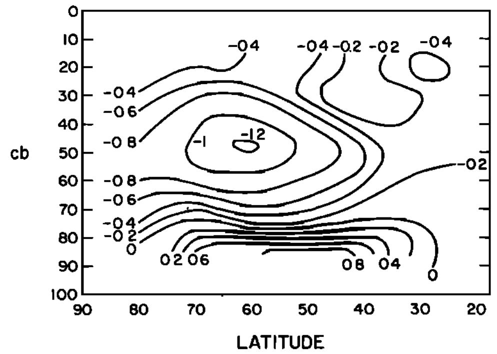

Fig. 23. Winter data (averages for January, February, March, October, November, and December 1963) of the mean meridional transport of quasi-geostrophic potential vorticity as a function of latitude and pressure in units of 10^{-4} ms^{-2} [after *Wiin-Nielsen and Sela*, 1971].

has been difficult to describe physical processes with the same detailed time and space resolution needed to view the daily weather. Consequently, most models of upper-atmosphere processes are in a sense climate models. The GCM's have attempted simulations up to 36 km. Simpler models have been developed to describe the coupling between motions, temperature, radiation, and composition at higher levels, e.g., the stratosphere (up to 55 km), mesosphere (55–85 km), lower thermosphere (85–120 km), and thermosphere (above 120 km). Some specific topics either of special current interest or germane to modeling the lower-atmosphere climate are summarized very briefly below.

First, there is the question of the distribution of stratospheric ozone, its dependence on tropospheric sources, and possible perturbations of its distributions from supersonic aircraft pollutants. (The U.S. Department of Transportation (DOT) is engaged in a large-scale study of this and other issues connected with the environmental effects of high flying aircraft. The DOT is supervising the preparation of a series of monographs summarizing present knowledge of all aspects of this question ranging from the engine effluents to the economic consequences of possible environmental changes.) The possible role of nitric oxides (NO and NO_2) as the dominant catalyst for recombination of ozone with atomic oxygen was recently recognized by *Crutzen* [1970, 1971] and *Johnston* [1971].

Ozone is responsible for shielding the surface environment from biologically hazardous ultraviolet (UV) radiation. Variations in UV radiation (associated with changes in the column content of ozone above a given location) can be related to, for example, variations in the incidence of skin cancer. Furthermore, stratospheric temperatures are maintained at their observed climatological values primarily by the balance between absorbed incoming solar UV radiation and emitted outgoing thermal radiation. Consequently, changes in the ozone distribution are followed by changes in the stratospheric temperature; for example, *McElroy et al.* [1974] argue that a reduction of the order of 30% in ozone concentrations in the stratosphere (by the NO emitted by a hypothetical fleet of supersonic transport planes) is accompanied by a 5°–10°K decrease in temperature in the stratosphere above 30 km (no calculation was made below 30 km). The chemical reactions important for stratospheric ozone are now reasonably well understood, but observational description of the global distribution of various trace constituents that affect ozone photochemistry is still inadequate. Models that can realistically simulate the transport of trace constituents on the time scale of years are essential for accurate prediction of the effect of stratospheric perturbations, but adequate versions of these models are not yet available.

The stratosphere couples to the troposphere through radiation and dynamic processes. For example, the radiative convective model of Manabe and Wetherald shows that changes in stratospheric temperature (because of perturbations in the concentration of water vapor or carbon dioxide (CO_2) concentrations) are generally accompanied by smaller (about one-fifth as large) tropospheric temperature changes of opposite sign [*Schneider and Coakley,* 1974]. Less ozone implies that UV radiation penetrates to and heats lower levels, implying a cooler stratosphere as found by *McElroy et al.* [1974] and thus increasing the net downward radiation reaching the surface. With increased CO_2 the tropospheric temperature change is dominated by the increased trapping of upward thermal radiation from the ground, whereas the stratospheric temperature change is dominated by CO_2 infrared cooling to space.

Various dynamic phenomena in the stratosphere are currently being modeled and related to tropospheric coupling. These phenomena include tropical wave modes and the 'quasi-biennial' oscillation (for example, see the recent review of *Wallace* [1973]), winter hemisphere stationary planetary waves [e.g., *Matsuno,* 1970], and 'sudden warmings' [e.g., *Matsuno,* 1971]. Some of these processes are now being simulated with GCM's [e.g., *Kasahara et al.,* 1973; *Manabe and Terpstra,* 1974].

In the mesosphere there has been great interest in exploring the connections between 'winter anomalies' in *D* region ionization and chemical transports related to planetary waves originating in the lower atmosphere [e.g., *Geller and Sechrist,* 1971].

One of the more fundamental questions regarding the lower thermosphere has been what processes are most important for determining the transition from heterosphere (gases being fully mixed) to homosphere (each gas being stratified according to its own hydrostatic balance) and in particular, what determines the relative latitudinal and time variations of various major gases, i.e., N_2, O_2, and O [e.g., *Johnson and Gottlieb,* 1973; *Hays et al.,* 1973].

Our understanding of the structure and dynamics of the thermosphere is now rapidly increasing, in part as a consequence of many new measurements. Important aspects of thermospheric climate include interaction of neutral gas with the ionosphere and the spatial and temporal variations of relative concentrations of major constituents. Recent reviews of currently significant problems have been given by *Dickinson* [1972*a*] and *Dickinson and Rishbeth* [1973].

It is important to understand better the climate of the upper atmosphere, not only because these regions are used directly for aerospace and other human activities but also because coupling of the tropospheric climate to variations of solar UV radiation and particle emissions must necessarily involve an upper-atmosphere linkage.

b. Planetary atmospheres. The planetary exploration program of NASA has stimulated intense interest in the climates of Mars, Venus, and Jupiter. Mars is characterized by a CO_2 atmosphere of the order of only 5-mbar surface pressure, a rotation comparable to that of the earth, and a solar constant reduced by approximately a factor of 2 in comparison with that of the earth. Mariner 8 and 9 reached Mars in the midst of a massive dust storm, and much effort has been devoted to interpreting this event [e.g., *Leovy et al.,* 1973]. The dust appears to have influenced drastically the global radiation balance, producing a near-isothermal atmosphere many tens of degrees warmer than is possible when CO_2 is the only radiatively active constituent [*Hanel et al.,* 1972]. It is hoped that an understanding of the radiative effects of dust in the Martian atmosphere will provide insights into the climatic importance of terrestrial dust. Also present on Mars (in exaggerated form compared to their occurrence on earth) are global scale variations in topography, and studies of their influence on Martian climate are expected to have spin-off for terrestrial climate modeling.

The third feature distinctive to Mars is an atmosphere consisting of a condensable volatile (i.e., CO_2), which may be frozen out and stored in the polar caps. *Gierasch and Toon* [1973] and *Sagan et al.* [1973] have discussed the possibility of consequent large variations in atmospheric mass and hence drastic changes from the present Martian climate. They point out the strong positive feedback between poleward transport of heat and polar condensation or evaporation. On the one hand, the Martian atmospheric pressure is equal to the vapor pressure (for CO_2) at the solid phase (i.e., at the polar 'ice' cap), which in turn is very sensitive to temperature. On the other hand, poleward heat transports are proportional to atmospheric mass. Thus if the atmosphere were warmed, much more heat would be exported from the Martian tropics to polar regions, resulting in increased surface pressures due to the temperature-dependent sublimation of the polar caps. *Sagan et al.* [1973] argue that this positive feedback (see also our section B2) could have increased the surface temperature (by an increased greenhouse effect) sufficiently to create water flows on the Martian surface that in turn carved the spectacular topography evidenced by Mariner 8 and 9.

Venus is characterized by a very slow retrograde rotation (243 days) and a long solar day (equal to 118 earth days). Processes at cloud level (around 200-mbar pressure and 240°K temperature) have been studied extensively from earth. Most of the thermal radiation escaping to space originates from cloud tops. The visible clouds are identified from polarization measurements as spherical droplets of 1.0-μm radius and refractive index of 1.44, which suggest their composition to be approximately a combination of four parts H_2SO_4 to one part H_2O [*Young*, 1973; *Hansen and Hovenier*, 1974]. They produce a planetary albedo of 0.77. Displacements and Doppler shift measurements of clouds around the 30-mbar level indicate a zonal retrograde motion of around 100 m/s [e.g., *Scott and Reese*, 1972]. Venera probes of the USSR have reached the planetary surface, measuring a near-adiabatic lapse rate below cloud level to the surface, where there is a pressure of 90 atm (earth) and a temperature of 750°K.

Several finite difference dynamic models for the climate of the Venusian troposphere have recently been integrated [e.g., *Sasamori*, 1971; *de Rivas*, 1973]. *Stone* [1974*b*] has developed a simple 'scale analysis' model for the mean horizontal and vertical temperature gradients of the Venusian troposphere and has shown that the combined effects of a Hadley-type circulation will always increase the stability of the vertical stratification. Thus a radiation greenhouse effect is required to explain the near-adiabatic lapse rates. Radiatively unstable vertical temperatures are shown to reach a slightly stable equilibrium temperature profile through dynamic processes on the 'advective' time scale, whereas with a stable radiative equilibrium profile steady state is reached only on the radiative time scale and is only slightly modified by the dynamics. The role of solar heating in providing a momentum source for the '4-day circulation' at cloud level has been treated by several authors, most recently by *Fels and Lindzen* [1974] (wherein earlier references may be found).

The simplest approach to vertical temperature structure is that of a radiative-convective equilibrium model. *Gierasch and Goody* [1970], for example, have developed such models for the Venusian troposphere but believed it, however, impossible to calculate jointly a radiative-convective equilibrium in which clouds are included that is also in agreement with optical and spectroscopic observations of the cloud properties. Above cloud level the global mean temperature profile shows a high degree of stable stratification, so that the global mean temperature structure above cloud level should be determined by a balance between absorbed solar radiation and infrared emission (i.e., radiative processes dominate convection in this region). Such a radiative model, assuming pure CO_2 composition, has been developed by *Dickinson* [1972*b*], including a thorough treatment of the large number of CO_2 infrared bands important for emission or absorption. The model is extended into the thermosphere, where molecular conduction also becomes important. For a recent review of other Venusian studies see the paper by *Marov* [1972].

Jupiter rotates more than twice as fast as earth and is a factor of 10 greater in radius. Of special interest have been the banded cloud structure and the equatorial zonal jet. *Williams and Robinson* [1973] have simulated both of these features in a numerical model for convection in a rotating spherical shell. *Starr* [1973] has discussed relationships between terrestrial and Jovian dynamic climatology.

For the largest (i.e., cosmogonical) time scales it becomes interesting to make comparative studies of the chemical evolution of the different planets. It has been argued by *Ingersoll* [1969] and *Rasool and de Bergh* [1970] that earth and Venus started under very similar conditions, both planets outgassing CO_2 and H_2O from the interior. On earth the CO_2 became locked up in rocks, and the H_2O stored in oceans, whereas on Venus the H_2O and CO_2 apparently stayed in the atmosphere, most of the H_2O being dissociated by UV radiation and the resulting hydrogen being lost to space. Rasool and de Bergh conclude that the runaway greenhouse effect in the Venusian atmosphere's evolution could have occurred on earth if its orbit were only 6–10 million km closer to the sun and would have made conditions on earth as hostile to life as those on Venus.

We hope that our necessarily brief discussion of the study of planetary atmospheres has left the impression that this not only is a subject of considerable intellectual interest but also has value for the study of the earth's atmosphere, since all planetary atmospheres obey the same basic physical laws but each exhibits unique characteristics that make intercomparison of different atmospheres an instructive exercise.

2. *Future of Climate Modeling*

This paper has dealt with some fundamental questions of climatic predictability, signal, noise, and equilibrium, the ingredients of a theory of climate, the general methodology of climate modeling, and some detailed examples selected from the hierarchy of climate models. If we have left the impression that the state of the art of climate theory and modeling is somewhat fragmented, we have merely reflected our view of the situation at present. Yet we hope our arguments prove that even today we possess much of the scientific information required to construct models of our present climate and to identify the dominant physical processes of climate change. Such climate models will be indispensable tools for the understanding and ultimate prediction of climate change, but these tools are as yet unsharpened, and there are many alternative modeling approaches, none yet being clearly superior to the others.

Such diverse disciplines as mathematics, hydrology, glaciology, biology, and chemistry play a role in the structure of various models of climate, so that improved interdisciplinary communication and cooperation will be essential for progress in the understanding of climate change. This communication must also go beyond scientific disciplines per se, for there are very real and vital questions to be addressed concerning the world's responses to any newfound ability to predict the climate. Physical science researchers developing the tools should not isolate themselves from their social and political scientist colleagues who are dealing with the social consequences of the new technology (for example, as is discussed by *Schneider* [1974*b*]).

Returning to questions of the future directions of climate modeling, we reiterate our feeling that a multiplicity of approaches must be tried and coordinated.

On the face of it the latest versions of the high-resolution GCM's, which (in their most general form) treat the dynamics of the oceans as well as the atmosphere, appear to be natural and essential tools in the study of climate and climate change. As we have described, the most recent models simulate the mutual interactions of the dynamic processes of the atmosphere with processes of radiative transfer, convection, and 'moist' thermodynamical processes, transport through the atmospheric boundary layer and subsurface layers, and important elements of the dynamics of the oceans. Improved variants of existing GCM's will also include the effects of variable gaseous species and aerosols.

It is evident, however, that detailed mathematical simulation of the three-dimensional behavior of the atmosphere cannot be the sole avenue to the study of long-period variations of climate (e.g., periods of the order of a decade or longer) and that the structure of the current or near-future GCM's may be unsuitable for that purpose on several counts:

1. The computing requirements for detailed climate simulation with current GCM's are prohibitive and will be prohibitive in the foreseeable future. Even one of the next generation of computers would take about a year to calculate a century of climate for a single combination of initial conditions or externally prescribed parameters. Moreover, to be of real value for climate studies, it may be necessary to calculate statistics from a finite ensemble of numerical integrations.

2. The GCM's have deliberately been designed to treat the dynamics of the atmosphere with great fidelity, whereas interactions of the atmosphere with the sun and underlying land or ocean surfaces have been treated relatively crudely. This approach is appropriate, of course, if the main concern is to predict the behavior of the atmosphere over short periods of time—i.e., periods short enough that it is not necessary to predict subtle changes in the characteristics of the underlying surface. On the other hand, it is plausible to assume that the dominant factors in climate change are the factors that have changed most, namely, the characteristics of the land and sea surfaces—reflectivity, conductivity, heat capacity, phase, etc.

3. There is an enormous disparity between the time scale of atmospheric variations and that of changes in the characteristics of the underlying land or sea surfaces.

4. Interpretation of cause and effect linkages may be difficult to trace in a GCM because of the large number of internal degrees of freedom in the model and because of the huge volume of output generated by a high-resolution time-dependent model.

Nonetheless, the essentially three-dimensional nature of the climate may require a model akin to the current GCM's for useful simulations. However, instead of following the present-day trend toward increased spatial resolution (primarily to improve the simulation of eddy flux transports) perhaps we should, at least for the sake of economy as is spelled out above, consider coarsening the resolution and concentrating on improving the eddy flux transport statistics through judicious choices of parameterization techniques. With the advent of the next generation of computing hardware such a limited resolution three-dimensional statistical-dynamic model of the atmosphere could be integrated routinely over century-long periods, permitting us greater flexibility and depth in modeling the relatively slowly varying conditions at the air-sea or air-ice interface.

Although it is clear that the existing or near-future high-resolution GCM's are not the only (or perhaps even the primary) weapons for attacking the problem of predicting long-term climatic variations, they play an indispensable role in testing and calibrating any statistical-dynamic models of climate and indeed in guiding the design of such models. Also, GCM's appear to be the only available means to include simultaneously the details of most of the important processes that affect climate. The GCM's must become a surrogate laboratory for the climate system itself. For example, lacking a sufficiently detailed and complete description to document climate changes over periods of several decades to centuries, we must test the detailed behavior of lower-order (especially statistical-dynamic) models against a limited number of GCM runs. The other essential step in the verification of each kind of model is to check both the detailed behavior of the GCM's and their statistical characteristics against the observed behavior of the atmosphere. It should be added that the time scale of climate changes between the unpredictability period of the atmosphere (a few weeks) and a few years may be the most socially and economically important period for which climatic prediction should be available (e.g., for crop-planting strategies to help alleviate food shortages). Study of the seasonal or interannual time scale of climate changes is tractable with high-resolution GCM's, and these models may play an indispensable role in investigating such variations.

Solid progress toward an understanding of the dominant factors in climate change will require steady development of an almost continuous spectrum or hierarchy of models of increasing physical and mathematical complexity. These models range, as we have shown, from simple (figuratively, 'back of the envelope') calculations of the effects of isolated processes on average conditions over the whole globe to space- and time-dependent computer simulations of a highly interactive system containing all known important processes.

The use of very simple models is primarily that of making tentative estimates of the sensitivity of long-term conditions of the atmosphere-land-ocean-cryosphere system to changes in various known thermodynamical and transport processes with a view to identifying those of greatest importance. This step is clearly essential to the design of more complicated interactive models, which in turn are equally essential to the design and interpretation of the behavior of still more realistic models.

Throughout the successive stages of model development, purely descriptive and diagnostic studies will play a key role: first, in suggesting simplifications and empirical parameterizations in the models and second, in providing vital verification and 'calibration' checks. Thus there must be a continual interplay between (1) the observed statistics of the real atmosphere on many time scales, (2) the derived statistics of GCM's, and (3) the results of statistical-dynamic models of climate and other models of lesser complexity. (Perhaps it might be useful here to include a further note of caution on the limitations of simple models for climatic predictions. J. Smagorinsky (private communication, 1974) has expressed this warning clearly: 'climatic variation is a result of very subtle and delicate interactions within the total atmosphere-ocean-continent-cryosphere system. This puts a burden of devising a parameterization of the synoptic scale dynamics compatible with the needed physical sophistication. Admittedly, many of the simple climatic models already developed have very great didactic value, but the critical question that remains is can one construct such models with sufficient mechanistic similitude in all other known aspects to provide reliable climatic predictors? This might not necessarily be the case if climatic change were essentially due to clearly external factors such as orbital intricacies. It may very well be that entirely different methods for reducing the computational burden will have to be found in which the full physical complexity can be retained as needed.' A further note from M. I. Budyko (private communication, 1974) should also be given in this context: 'Though at present we are far from distinct understanding of the degree of detail in numerical models essential for solving different climatological problems, it is evident that some aspects of climate genesis might be elucidated by means of the simplest existing models, while for studying many questions of climate theory the most general models presently available are insufficient.')

In conclusion, climate modeling has possibly now reached a threshold where further progress will lead to potential human benefits.

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