Annu. Rev. Fluid Mech. 1994. 26: 353–78

CLIMATE DYNAMICS AND GLOBAL CHANGE

R. S. Lindzen

Center for Meteorology and Physical Oceanography, MIT, Cambridge, Massachusetts 02139

KEY WORDS: climate variability, greenhouse effect, glacial cycles, carbon dioxide, water vapor

1. INTRODUCTION

The question of global climate change has been a major item on the political agenda for several years now. Politically, the main concern has been the impact of anthropogenic increases in minor greenhouse gases (the major greenhouse gas is water vapor). The question is also an interesting scientific question. Restricting ourselves to issues of climate, the answer requires that we be able to answer at least two far more fundamental questions, both involving strong fluid mechanical components:

1. What determines the mean temperature of the Earth?; and
2. What determines the equator-pole temperature distribution of the Earth's surface?

The popular literature lays stress on the first question, but the two are intimately related, and there are reasons for considering the second question to be the more fundamental. When we discuss climate observations, we will see that climate changes in the past history of the Earth were primarily associated with almost unchanged equatorial temperatures and major changes in the equator-pole distribution. There are good reasons to view changes in the mean temperature of the Earth as residual terms arising from the change in the equator-pole distribution. Section 2 of this review quickly summarizes observations of climate. We discuss not only the temperature trends of the past century, but also earlier climate. In

0066–4189/94/0115–0353$05.00

connection with the discussion of glaciation cycles, we also introduce the Milankovitch hypothesis, which attempts to relate these cycles to variations in the Earth's orbital parameters. Section 3 reviews our current understanding of the two fundamental questions. The answer to both questions depends on the heat budget of the Earth—both its radiative and dynamic components. The radiative contributions depend on the radiatively active constituents of the atmosphere—mainly water vapor and cloud cover. In the context of the present political debate, the focus is on the minor greenhouse gases (primarily CO_2), and the behavior of water vapor and clouds is subsumed under the title "feedbacks." In present models, water vapor automatically increases with warming, and constitutes the major positive feedback. Without this feedback, no current model would produce *equilibrium* warming due to a doubling of CO_2 in excess of about 1.5°C regardless of other model feedbacks. This brings us to two additional questions: Namely, what determines the density of water vapor in the atmosphere?, and how is the time-dependent response of the atmosphere to radiative perturbations related to the equilibrium response? Sections 4 and 5 deal with these two questions.

The time-dependent behavior of the climate is highly contingent on the presence of oceans. This is true not only for temperature, but also for CO_2. The behavior of CO_2 strongly involves chemistry and largely transcends the scope of this review. However, we briefly discuss this issue in Section 6. Section 7 summarizes our discussion, emphasizing relatively simple and focused approaches to the question of how we might expect the climate to respond to increased emissions of CO_2. At present, we note that there is no basis in the data for current fears, and model predictions result from physically inadequate model features. We suggest reasons for expecting small warming to result from expected increases in CO_2.

2. OBSERVATIONS OF CLIMATE

The following is only a cursory treatment of the observed climate, focusing only on those aspects essential to subsequent discussion. Houghton et al (1990, 1992), Crowley & North (1991), Imbrie & Imbrie (1980), Balling (1992), and Peixoto & Oort (1992) provide much more material for those interested.

The definition of climate variations is not without ambiguities. We shall take climate variability to refer to changes on time scales of a year or longer. In this paper, moreover, we will restrict ourselves to surface climate on a global scale. For simplicity, most global change studies have focused on the globally and annually averaged temperature. The behavior of this quantity since 1860 is illustrated in Figure 1. Figure 1 is based primarily on temperature records over land. Attempts have been made to use primarily records from areas minimally affected by urbanization, though urban heat island effects may have introduced errors on the order of 0.1°C. Ocean data are limited, and such data as are available have been "corrected" substantially (several tenths of a degree). In forming global averages, data have been interpolated over a regular grid enabling one to take area-weighted averages. Given the preponderance of data from Northern Hemisphere land stations, this means that large areas are based on minimal observation. What Figure 1 shows is a global temperature that rose noticeably between 1915 and 1940, remained relatively steady until the early 1970s, and rose again in the late 1970s. The change over the past century is estimated to be $0.45°C \pm 0.15°C$. The temperature increase in the late 1970s is largely a Southern Hemisphere phenomenon. In the Northern Hemisphere, there was a decline in temperature from the mid 1950s until the early 1970s. A more complete discussion and extensive references may be found in Houghton et al (1990) and Balling (1992). It is clear that there has been a leveling of temperatures in the 1980s, and although these are referred to as record-breaking years, they are not appreciably warmer than the record-breaking years of the 1940s. Recently, satellite data have been used to obtain global average temperature (Spencer & Christy 1990). Such data are available since 1979. The data tend to be representative of the whole troposphere rather than the surface. According to all existing models, climate response to additional greenhouse gases should affect the entire troposphere, so that the satellite data might, in some respects, be preferable to surface data. On the whole, satellite data correlate well with surface data when the latter are plentiful, and relatively poorly when surface data are sparse. Trenberth et al (1992) have recently reviewed this situation, concluding that surface data from the late 19th century may be less reliable than claimed. However, even without this caveat, the Intergovernmental Panel on Climate Change (Houghton et al 1990) notes that the surface record depicts nothing that can be distinguished from natural variability.

Proxy records exist that offer some suggestion of how climate varied prior to the instrumental record. However, these records are usually insufficient for global averages. There is some evidence for a "little ice age" in the 18th Century, as well as a medieval optimum when temperatures were significantly warmer than at present (Crowley & North 1991). Budyko & Izrael (1991) in reviewing past climates notably different from the present observed that these climates differed from the present not only in mean temperature but in temperature distribution with latitude. They argue for a universal distribution in latitude shown in Figure 2. This distribution is characterized by very small changes near the equator, and

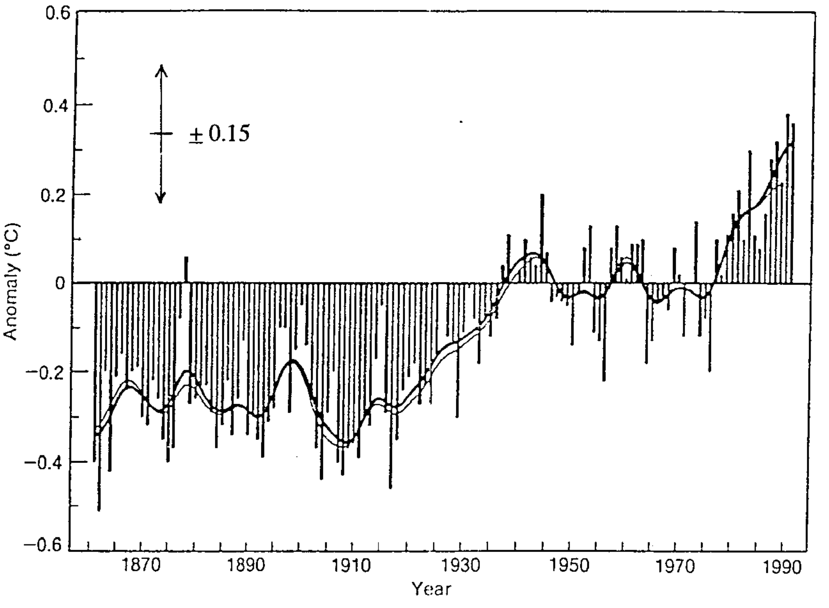

Figure 1 Globally averaged surface temperature record since 1860. (From Houghton et al 1992. Light line from Houghton et al 1990.)

major changes in the equator-to-pole temperature difference. This hypothesized "universal" curve poses two major questions:

1. Since the change of equator-to-pole temperature difference must indicate a change in the heat flux from the tropics to higher latitudes, why are the variations at low and high latitudes not out of phase with each other?
2. What is preventing significant variation of equatorial temperatures? The crucial point here is that the changing equator-to-pole temperature differences would appear to call for profound changes in the heat flux out of the tropics, which for the tropics represents a large change in thermal forcing.

In this connection, it should be noted that there are suggestions (Barron 1987) that during the very warm climate of the Eocene (~ 50 million years ago) the equator may have been colder than at present. Also, in connection with the "warming trend" of the past century, the pronounced latitude variation of Figure 2 was absent, and in the warming episode of the late 1970s, tropical warming exceeded polar warming, which may even have amounted to cooling.

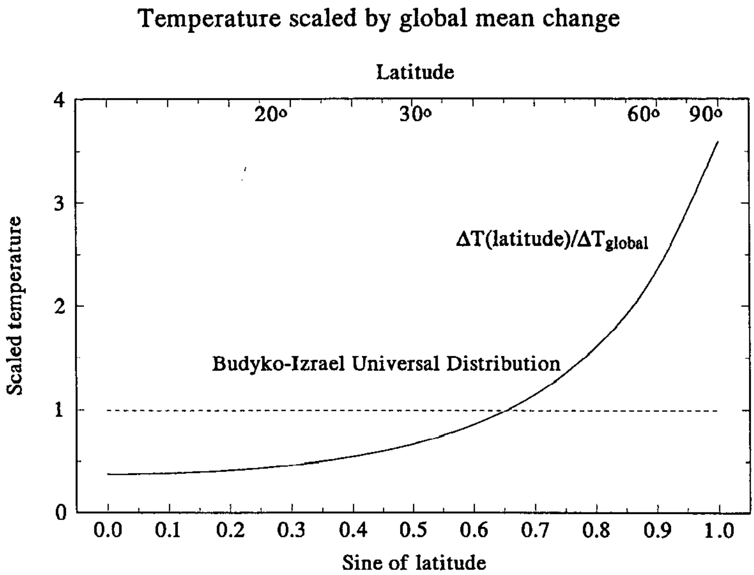

Figure 2 Universal latitude variation of climate change. (After Budyko & Izrael 1991.)

The past million years or so have also manifested an additional striking aspect of climate: cycles of major glaciation and deglaciation. The cycles are determined from the study of δO^{18} in ice cores. δO^{18} is primarily indicative of ice volume. While there are problems in dating different levels in such cores, Figure 3 gives a widely accepted time history from such cores. Figure 4 shows a power spectrum of this time series. The 100,000 year component clearly dominates, but significant peaks are claimed near

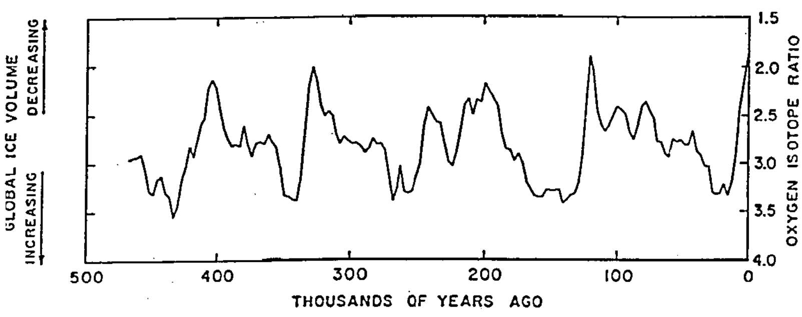

Figure 3 δO^{18} as a function of time over the past 700,000 years. (From Imbrie & Imbrie 1980.)

40,000 and 20,000 years. It should be noted that there remain arguments concerning these peaks based on both core dating (Winograd et al 1992) and analysis method (Evans & Freeland 1977). The general hypothesis for this periodic behavior is that the climate has been forced by the orbital variations of the Earth; this is referred to as the Milankovitch mechanism. The orbital variations consist of variations in the obliquity (tilt) of the Earth's rotation axis, the precession of the equinoxes along the Earth's elliptic orbit, and the changes in eccentricity of the orbit. These variations are schematically illustrated in Figure 5. The obliquity variations are characterized by periods of about 40,000 years, the precession is associated with periods of about 20,000 years, and the eccentricity is associated with periods of 100,000 years and 400,000 years. The precessional cycle, moreover, is strongly modulated by the eccentricity cycle (Berger 1978). The relevant periods are indicated in Figure 4. Both the 24,000 yr and 19,000 yr peaks correspond to the precessional cycle. Imbrie & Imbrie (1980) provide a readable treatment of this phenomenon. From the point of view of climate change, there are several aspects of the glaciation cycles

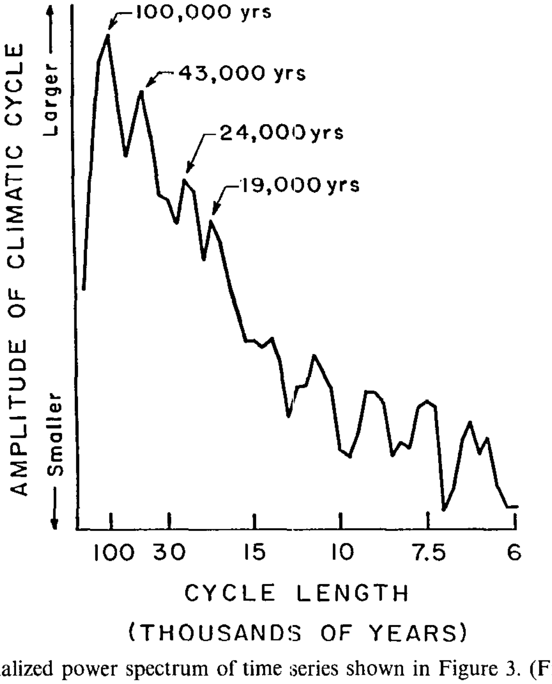

Figure 4 Normalized power spectrum of time series shown in Figure 3. (From Imbrie & Imbrie 1980.)

that deserve comment. First, the change in annually and globally averaged insolation associated with orbital variations is very small ($\leqslant 1\%$). On the other hand, orbital changes lead to substantial changes in the geographical distribution of insolation. Milankovitch (1930) stressed the importance of summer insolation at high latitudes for the melting of winter snow accumulation. More recently, Lindzen & Pan (1993) have noted that orbital variations can greatly influence the intensity of the Hadley circulation, a basic component of planetary heat transport. Relatively uniform changes in heating, it should be recalled, will have little effect on heat transport.

3. BASIC PHYSICS OF GLOBAL CLIMATE

Global Mean Temperature

The most common but, as we shall see, severely incomplete approach to global mean temperatures is to consider a one-dimensional radiative convective model with solar insolation characteristic of some "mean" latitude at equinoxes. An example of such an approach is illustrated in Figure 6, taken from Möller & Manabe (1961). Several vertical profiles of temperature are shown: pure radiative equilibrium, with and without the infrared properties of clouds, and radiative-convective equilibrium with the infrared properties of clouds included. (The visible reflectivity of clouds is included in all the calculations.) In most popular depictions of the

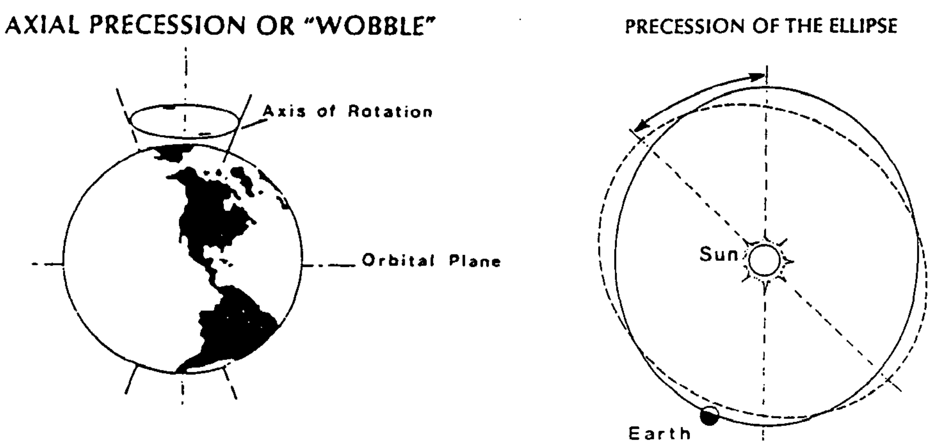

Figure 5 Schematic illustration of orbital parameters involved in Milankovitch mechanism. (After Crowley & North 1991.)

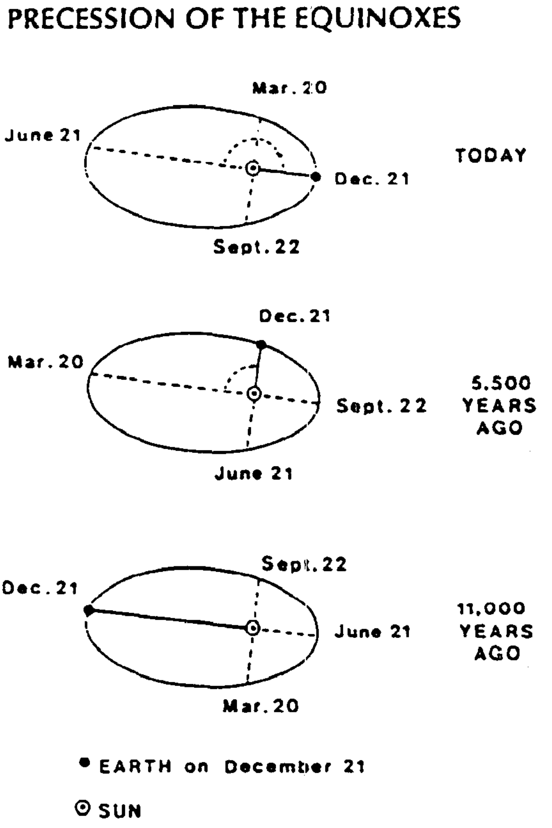

Figure 5—(*continued*)

greenhouse effect, it is noted that in the absence of greenhouse gases, the Earth's mean temperature would be 255 K, and that the presence of infrared absorbing gases elevates this to 288 K. In order to illustrate this, only radiative heat transfer is included in the schematic illustrations of the effect (Houghton et al 1990, 1992); this lends an artificial inevitability to the picture. Several points should be made concerning this picture:

1. The most important greenhouse gas is water vapor, and the next most important greenhouse substance consists in clouds; CO_2 is a distant third (Goody & Yung 1989).
2. In considering an atmosphere without greenhouse substances (in order to get 255 K), clouds are retained for their visible reflectivity while ignored for their infrared properties. More logically, one might assume

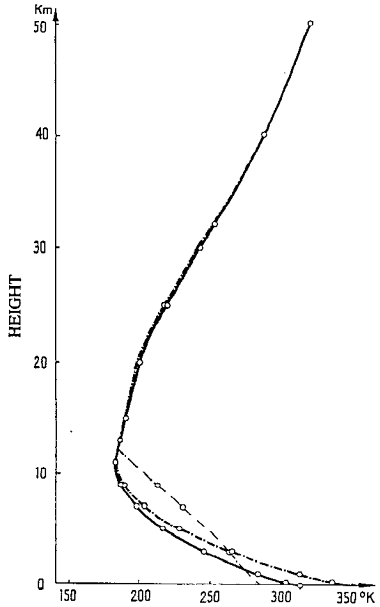

Figure 6 Pure radiative equilibrium with infrared effects of clouds included (*thick dashed-dotted curve*) and without clouds (*solid curve*); radiative-convective equilibrium (*thin dashed curve*). (After Möller & Manabe 1961.)

that the elimination of water would also lead to the absence of clouds, leading to a temperature of about 274 K rather than 255 K.

3. Pure radiative heat transfer leads to a surface temperature of about 350 K rather than 288 K. The latter temperature is only achieved by including a convective adjustment that consists simply in adjusting the vertical temperature gradient so as to avoid convective instability while maintaining a consistent radiative heat flux. [It should be noted that this is a crude and inadequate approach to the treatment of convection; however, the development of better approaches is still a matter of active research (Arakawa & Schubert 1974, Lindzen 1988, Geleyn et al 1982, Emanuel 1991).] The greenhouse effect can be measured in terms of the change in T^4 at the surface necessitated by the presence of infrared absorbing gases. From this perspective, the presence of convection diminishes the purely radiative greenhouse effect by 75%. The reason is that the surface of the Earth does not cool primarily by radiation. Rather, convection carries heat away from the surface, bypassing much of the greenhouse gases, and depositing heat at higher levels where there is less greenhouse gas to inhibit cooling to space.
4. Water vapor decreases much more rapidly with height than does mean air density. Crudely speaking, the scale height for water vapor is 2–3

km compared with 7 km for air. As a result of the convection in item 3 above, water vapor near the surface contributes little to greenhouse warming. A molecule of water at 10 km altitude is comparable in importance to 1000 molecules at 2–3 km, and far more important than 5000 molecules at the surface (Arking 1993). In fact, water vapor decreases rapidly not only with increasing altitude but also with increasing latitude. Thus, the mean temperature of the Earth will depend not only on vertical transport of heat but also on meridional transport of heat. In attempting to calculate the mean temperature of the Earth by means of one-dimensional models one is assuming that one can find a latitude where the divergence of the dynamic heat flux is zero. However, in the absence of knowledge of the horizontal transport the choice of such a latitude is no more than a tuning parameter.

The situation summarized in items 3 and 4 above is schematically illustrated in Figure 7.

The Equator-to-Pole Temperature Distribution

The current annually averaged equator-to-pole temperature difference is about 40°C. In the absence of dynamic transport this quantity would be about 100°C (Lindzen 1990). The difference is even more striking for the winter hemisphere where the polar regions do not receive any sunlight.

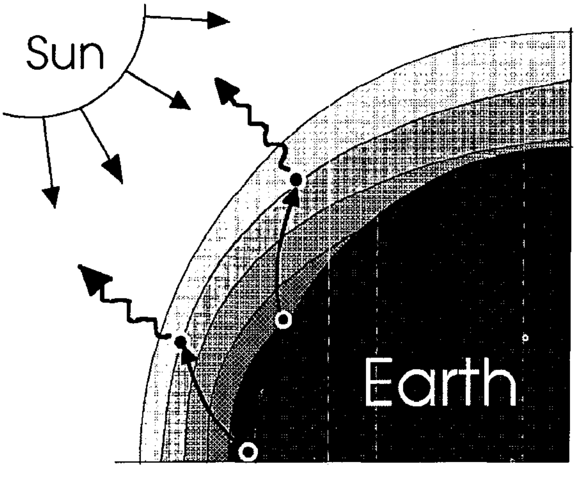

Figure 7 Schematic illustration of greenhouse effect with dynamic heat transfer. Infrared capacity is greatest at the ground over the tropics, and diminishes as one goes poleward. Air currents bodily carry heat to regions of diminished infrared opacity where the heat is radiated to space—balancing absorbed sunlight. Lighter shading schematically represents reduced opacity due to diminishing water vapor density.

Interestingly, there is currently no simple theory that quantitatively predicts the equator-to-pole temperature, despite the fact that it is this quantity that seems most relevant to climate change. As noted above, moreover, a knowledge of horizontal transport is also essential to calculating the global mean temperature. Current large-scale numerical models have difficulties here: both in the prediction of eddies (Stone & Risbey 1990) and in the prediction of polar temperatures (Boer et al 1992).

The usual picture is that heat is transported within the tropics by a large-scale cellular flow known as the Hadley circulation, and from 30° to the poles by baroclinically unstable eddies (Lorenz 1967). While this picture is roughly correct, it has a profound seasonal character. Except for a brief period when the zonally averaged surface temperature maximum is exactly at the equator, the Hadley circulation consists in a single cell with ascent in the summer hemisphere and descent in the winter hemisphere (Oort & Rasmussen 1970, Lindzen & Hou 1988). The descending branch typically extends 30° into the winter hemisphere and is associated with strong lateral gradients in potential vorticity (Hou & Lindzen 1992), and such gradients are generally associated with eddy instability. Indeed, eddy heat transport is much larger in the winter than in the summer (Oort 1983). The heat transport situation is schematically illustrated in Figure 8. The above only refers to the atmosphere; heat transport in the ocean is much less well understood, though it appears to be significant (Carrissimo et al 1985). Here we focus on the atmospheric transport for several reasons: the shallow ocean circulation is wind driven and is not a direct response to heating gradients; and the deep thermohaline circulation in the ocean is slow compared to the time it takes for the surface of the ocean to equilibrate thermally with the air above and the radiative forcing.

There is, in fact, reason to believe that the mean meridional temperature distribution is largely determined by the atmospheric transport. As noted above, heat transport between the tropics and high latitudes is carried by eddies that are believed to arise from baroclinic instability (Eady 1949, Charney 1947). The possibility that these eddies act to neutralize the basic state has long been suggested (Pocinki 1955, Stone 1978, Lindzen & Farrell 1980). However, the neutral states considered were based on the Charney-Stern condition (Charney & Stern 1962)—an extension of the Rayleigh inflection point condition to rotating stratified fluids (Lindzen 1990), and these states differed from the observed state rather profoundly. Recently, however, a new neutral state has been found (Lindzen 1993a) which is remarkably compatible with the observed state (Sun & Lindzen 1993c). The match between this state and the Hadley regime appears to depend on the intensity of the Hadley circulation (Hou 1993). The intensity of the Hadley circulation depends, in turn, on the displacement of the summer

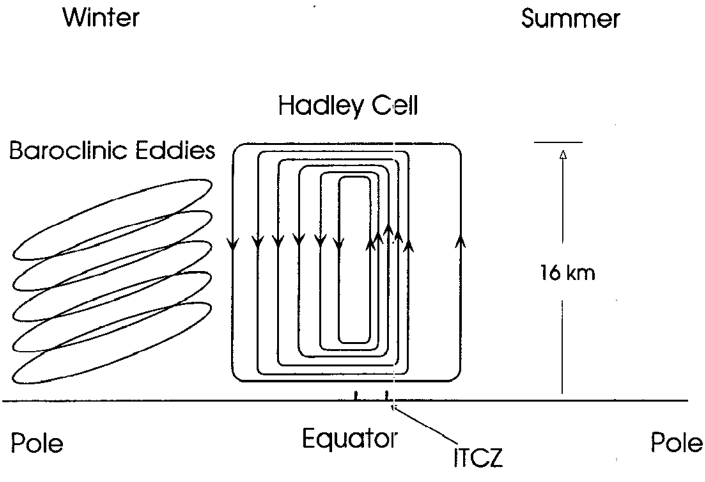

Figure 8 Schematic illustration of major dynamic mechanisms for meridional heat transfer in the atmosphere.

surface temperature maximum from the equator (Lindzen & Hou 1988) and on the sharpness of the maximum (Hou & Lindzen 1992). The former depends pronouncedly on the orbital variation (Lindzen & Pan 1993) and provides a possible physical basis for the Milankovitch mechanism. It thus appears that atmospheric processes alone may largely determine the gross temperature structure of the atmosphere (including the surface). This does not preclude an oceanic contribution to the heat flux—as long as it is not so great as to completely preclude atmospheric eddies.

4. CLIMATE SENSITIVITY AND FEEDBACKS

Current climate concerns focus on the response of the global mean temperature to increasing atmospheric CO_2. Sensitivity is defined, for convenience, as the equilibrium response of the global mean temperature to a doubling of CO_2. Such a definition is meaningful for the problem at hand, but in no way suggests that the sensitivity, so defined, is relevant to past climate change. As we have already noted, past climate change appears to have been primarily associated with changes in the equator-to-pole heat flux, and gross forcing such as that provided by increasing CO_2 does not significantly affect such fluxes in any way currently identified. Nevertheless, current large-scale numerical simulations of climate commonly suggest significant climate response to a doubling of CO_2, even in the tropics (Houghton et al 1990, 1992); there is no evidence in these models of any tropical stabilization. While these results immediately suggest a certain questionability concerning the models, it is of interest to examine the model response in terms of the physics they contain. It should first be noted that the expected globally averaged warming from a doubling of CO_2 alone without any feedbacks is 0.5–1.2°C (Lindzen 1993b). Model predictions of values from 1.5–4.5°C (Houghton et al 1990, 1992) depend on positive feedbacks within the models. One may write the globally averaged equilibrium warming for a doubling of CO_2 as

$$\Delta T_{2\times CO_2} = \text{gain} \times \Delta T_{ng}, \tag{1}$$

where ΔT_{ng} is the response to a doubling of CO_2 in the absence of feedbacks. Gain is related to feedback by the expression

$$\text{gain} = \frac{1}{1-f}, \tag{2}$$

where f is the feedback factor. To the extent that feedbacks from different processes are independent, their contributions to the feedback factor are additive: i.e.

$$f = \sum_i f_i. \tag{3}$$

Note that gains from various processes are not additive. Crude analyses have been made of the physical origins of feedbacks in various models.

Results for several commonly cited models are given in Table 1. As we see from Table 1, the largest feedback is from water vapor, and arises because in all models, water vapor at all levels increases with increasing surface temperature. Recall that it is upper-level (above 2–3 km) water vapor that is of primary importance for greenhouse warming. (Note that water vapor and lapse rate tend to be lumped together because lapse rate changes accompany water vapor changes in current models, and the lapse

Table 1 Feedback factors, f_i, for GISS[a] and GFDL[b]

Process	Model	
	GISS	GFDL
Water vapor/lapse rate	0.40	0.43
Cloud	0.22	0.11
Snow/albedo	0.09	0.16
TOTAL	0.71	0.70
gain [$= 1/(1-f)$]	3.44	3.33

[a] Goddard Institute for Space Studies.
[b] Geophysical Fluid Dynamics Laboratory at Princeton.

rate, itself, influences the surface temperature.) The next feedback arises from clouds. In most models, surface warming is accompanied by increasing cloud cover. It may seem surprising that this leads to a positive feedback, but in these models, the infrared properties of the clouds outweigh their visible reflectivity. The remaining feedback is from snow/albedo. This refers to the fact that in the models, increased surface temperature is associated with reduced snow cover, which in turn leads to reduced visible reflectivity. There is, in fact, substantial uncertainty over the last two feedback mechanisms (Cess et al 1990, Cess et al 1991). The cloud feedback may very well be negative rather than positive (Mitchell et al 1989, Ramanathan & Collins 1991). Even the snow/albedo feedback is subject to doubt because of such factors as winter-night, high-latitude cloud cover, etc. The magnitude of this feedback in existing models varies greatly from model to model. Oddly enough, there is a tendency to regard the water vapor/lapse rate feedback as well determined because most models behave similarly despite the fact that the physics relevant to the upper level water vapor is absent in these models. Upper level water vapor in these models appears to be determined by diffusion from below which, as we show later, is impossible. For the present, we should note that the water vapor feedback alone would only produce a gain of about 1.67, but its presence is essential to the total gain. Without it, no model would produce a ΔT_{2CO_2} greater than 1.7°C regardless of the presence of other feedbacks, and under the assumption that ΔT_{ng} has the large value of 1.2°C.

For our purposes, the central fact about water vapor is the Clausius-Clapeyron relation for the saturation vapor pressure. This quantity drops sharply with temperature as illustrated in Figure 9. A parcel of air suffers compressional heating as it descends, and adiabatic expansion and cooling as it rises (both at a rate $g/c_p = 9.8$°C/km). Thus a rising parcel that starts at 80% relative humidity (i.e. 80% of saturation) will saturate within a few hundred meters, while a saturated parcel at 12 km will have its relative humidity decrease exponentially (with a scale height of 2–3 km) with depth as it sinks. This situation is schematically illustrated in Figure 10. If the source of water vapor in the upper troposphere were water vapor deposited aloft by deep cumulus convection and carried downward, we would expect rapidly decreasing relative humidity with depth. Instead, relative humidity is relatively constant with depth. If the source were upward transport from below, we would expect widespread deep cloud up to the level being supplied. This too is not observed. Rather, it appears that the source of upper troposphere water vapor is the evaporation of precipitation originating in ice crystals near the tropopause (Smith 1992, Sun & Lindzen 1993b, Betts 1990). Such a process is free of the problems associated with

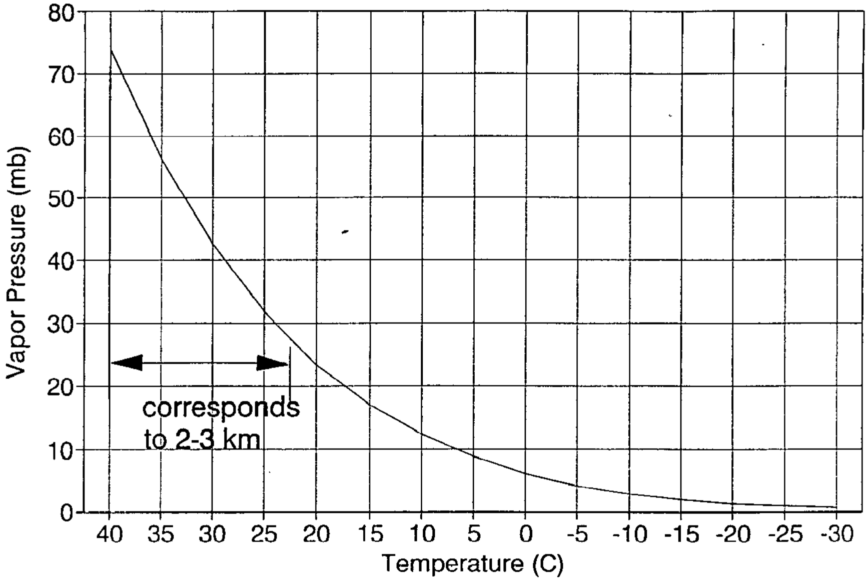

Figure 9 Saturation water vapor pressure (over pure liquid water) as a function of temperature.

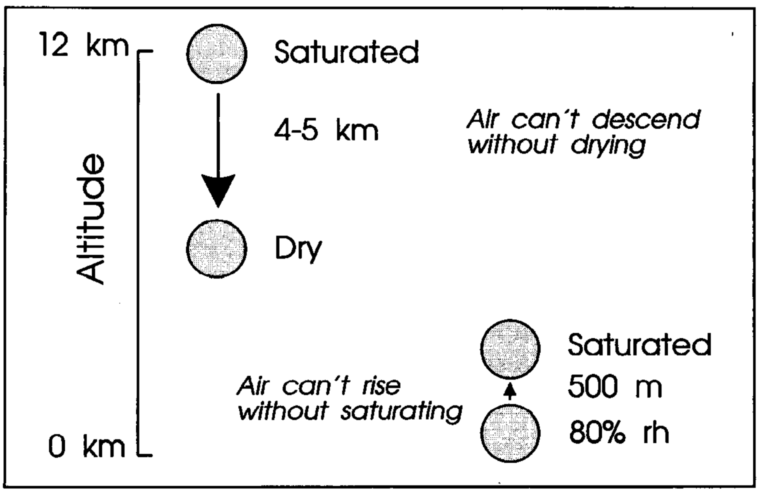

Figure 10 Schematic illustration of consequences of simple vertical transport of water vapor.

the direct transport of water vapor. In the tropics, at least, these crystals appear to have been directly detrained from cumulonimbus towers. What we need to know is how the amount of detrained ice is determined. The simplest models of cumulus convection (Arakawa & Schubert 1974, Lindzen 1988) assumed that as air rises in a cumulonimbus tower, all condensate rains out, leaving just saturated air to be detrained at the cloud top. At the very least, this requires slow ascent. With rapid ascent, at least some of the condensate will be lofted by the updraft to above the freezing level (ca 5 km) and detrained in the form of ice. The speed of the updrafts in cumulonimbus towers will be determined by the convective available potential energy (CAPE) of the ambient air. This simply refers to the integral of the buoyancy force over the depth of the cloud (Bolton 1980, Williams et al 1992). It seems reasonable to assume that the amount of detrained ice will depend on CAPE, and that the nature of the water vapor feedback will depend on how CAPE responds to warming.[1]

In this connection, it is important to note that the dependence of CAPE on surface temperature within a given climate regime may differ from the dependence of CAPE on global (or tropical) mean temperature in different climate regimes. The point is simply that (at least within the tropics), the large-scale circulation acts to eliminate horizontal temperature gradients above the trade wind boundary layer. However, within the turbulent boundary layer the horizontal temperature variation is less well mixed; warmer surface temperatures are therefore associated with greater CAPE. However, the possibility remains that in different climates, the contribution to CAPE from above the boundary layer may be different, and that warmer climates (as opposed to warmer local surface temperatures) may be associated with reduced overall CAPE (at least in the tropics). This was argued by Sun & Lindzen (1993a). Observationally, Oort (1993), analyzing routine radiosonde data for globally averaged temperature and specific humidity at various levels in the atmosphere, did find that the global warming of the late 1970s was indeed accompanied by reduced specific humidity above 700 mb. Indeed the observed reduction was such to produce a water vapor feedback contribution to f (in Equation 3) of -6 as opposed to $+0.4$ as is found in current models (Sun & Lindzen 1993a). Unfortunately, changes in sensors used to measure humidity make the water vapor time series unreliable (Elliot & Gaffen 1991).

Interestingly, however, the inferred value of $f = -6$ would be sufficient

[1] It should be noted that the supposition in a number of papers (Raval & Ramanathan 1989, Rind et al 1991) that upper level water vapor is determined simply by surface temperature immediately below is at variance with both the above physics, and the fact that over 99.9% of the air in the tropics is descending, not rising from the surface (Sarachik 1985).

to explain the stability of tropical temperatures despite changes in equator-to-pole heat fluxes equivalent to changes in surface fluxes of 10s of watts/m^2. It should be emphasized that the truly remarkable fact about tropical stability is its existence in the presence of very large changes in tropics-to-high latitude heat fluxes. These changes are far larger than those envisaged as being due to changing CO_2, solar radiation, etc.

5. EQUILIBRIA VS TIME-DEPENDENT RESPONSE TO CLIMATE PERTURBATIONS

The above discussion was framed in terms of the equilibrium response to climate perturbations. In point of fact, most of the Earth is ocean covered, and the heat capacity of the ocean will delay the response of the climate system as a whole. As noted by Hansen et al (1985), the delay will depend on the rate at which heat is transported downward in the ocean (the more rapidly heat is transported, the longer the delay), and the feedbacks in the atmosphere. The stronger the atmospheric feedbacks are, the longer the delay since, as we shall see, stronger atmospheric feedbacks imply weaker coupling to the sea surface. The simplest approach to ocean delay has been to use simple box-diffusion models for communication between the atmosphere and the ocean. A typical geometry is shown in Figure 11. The relevant equations for this geometry are:

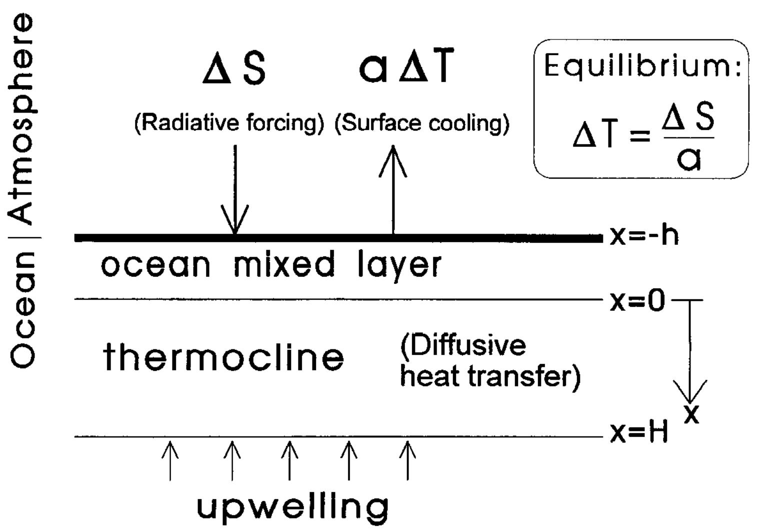

Figure 11 Geometry of simple box-diffusion model for ocean heat absorption. (From Lindzen 1993b.)

$$\rho ch \frac{\partial \Delta T_1}{\partial t} = \Delta S - a\Delta T_1 - \kappa \frac{\partial T_2}{\partial x}\bigg|_{x=0}, \tag{4a}$$

$$\frac{\partial T_2}{\partial x} = \kappa \frac{\partial^2 T_2}{\partial x^2} \quad \text{for} \quad x > 0, \tag{4b}$$

$$T_2 = T_1 \quad \text{at} \quad x = 0, \tag{4c}$$

$$\frac{\partial T_2}{\partial x} = 0 \quad \text{at} \quad x = H, \tag{4d}$$

where T_1 is the mixed layer temperature, and $T_2(x)$ is the temperature below the mixed layer, ρ is the density of water, c is the heat capacity of water, and h is the thickness of the mixed layer (taken to be 70 m). H is the depth of the thermocline below which upwelling inhibits downward heat diffusion. κ, the eddy heat diffusion, is typically taken to be 1.5 cm^2 sec^{-1}. ΔS represents a radiative perturbation, while $a\Delta T_1$ represents the response of the surface temperature. If $a = a_0$ in the absence of gain, then, in the presence of gain, $a = a_0/g$, and $g = 1/(1-f)$. As already noted the larger g is, the weaker the coupling in Equation 4a.

Equations 4 above are relatively standard equations for diffusive heat transfer. The response of the surface temperature to perturbed forcing is not simply exponential in time, but rather involves a continuously increasing time scale as heat penetrates to greater depths (Lindzen 1993). However, for purposes of discussion, it is convenient to define a characteristic time as that time over which the response to impulsive forcing reaches to within $1/e$ of its equilibrium value. A plot of how this characteristic time varies with gain is given in Figure 12. We see that the depen-

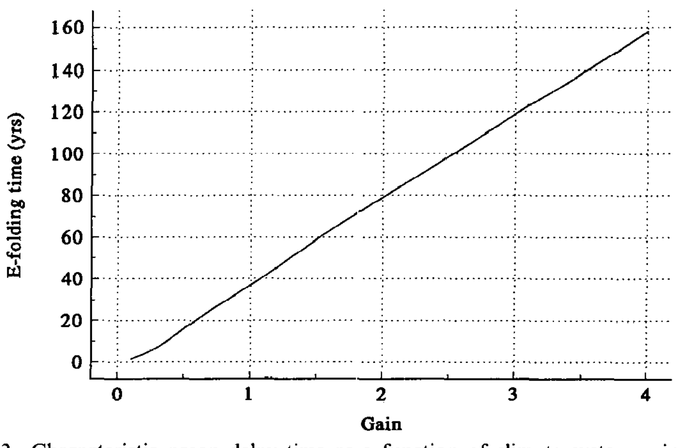

Figure 12 Characteristic ocean delay time as a function of climate system gain. (From Lindzen 1993b.)

dence is almost linear away from very small gains. This relation between climate sensitivity (i.e. gain) and ocean delay allows a more sophisticated approach to assessing both the past record of climate and projections for the future. In Figure 13, we show expected warming on the basis of various climate sensitivities for the IPCC (Intergovernmental Panel on Climate Change) business-as-usual emissions scenario (which involves quadrupling effective CO_2 by 2100). Although differing sensitivities lead to different results by 2100, the fact that higher sensitivities are associated with longer delays leads to the disappointing result that the record of the past century (when effective CO_2 increased by almost 50%) is broadly consistent with virtually any sensitivity (assuming natural variability on the order of 0.5°C). On the other hand, if we assume a gain of 3.33, and vary κ in order to obtain different delay times, we get the results shown in Figure 14. These results make it clear that such high gain is broadly consistent with the observed record only if the characteristic delay time is greater than 100 yrs.

A matter of some interest is whether one can directly measure delay time. For example, large volcanic eruptions provide a significant change in albedo for a year or so following eruption (Oliver 1976). The usual picture is that it takes about three months for volcanic emissions to spread around the world forming a sulfate aerosol layer which then decays with a time scale of about one year. The response of the system described by Equations 4 to such forcing (for different choices of gain) is shown in Figure 15. Given the numerous uncertainties associated with volcanic forcing, the responses during the first three years following an eruption do

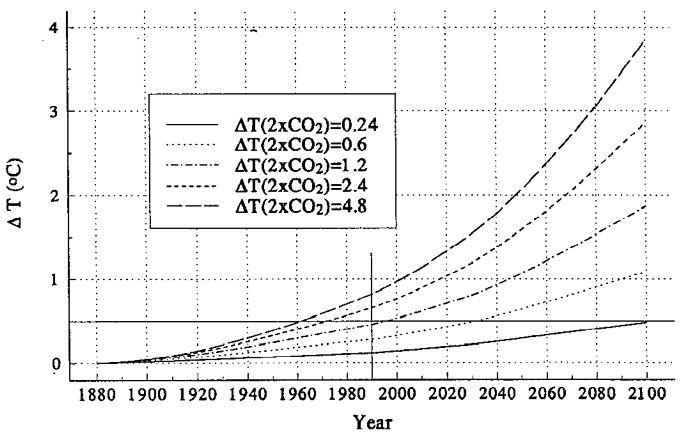

Figure 13 Change in global temperature expected for IPCC business-as-usual emissions scenario and various climate sensitivities. (From Lindzen 1993b.)

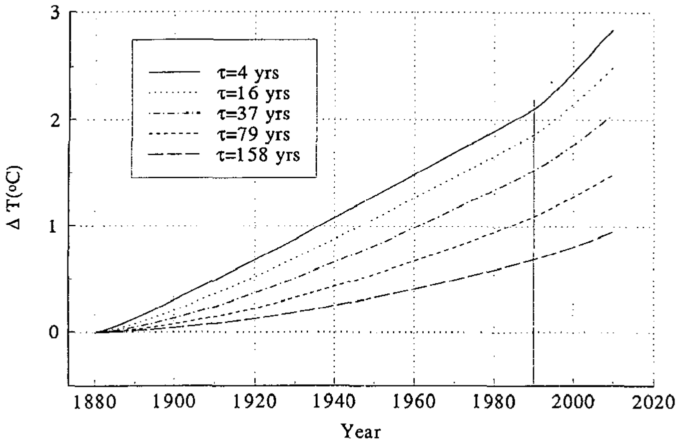

Figure 14 Behavior of global temperature for IPCC business-as-usual emissions scenario, $\Delta T_{2CO_2} = 4°C$, and various choices of ocean delay. (From Lindzen 1993b.)

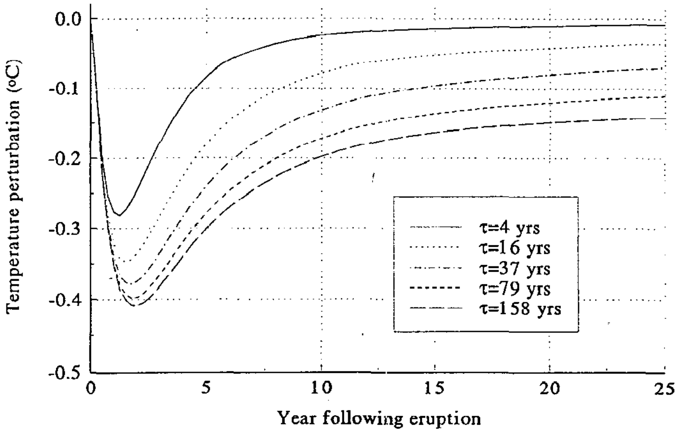

Figure 15 Response to Krakatoa type volcanic eruption for various choices of climate system gain. (From Lindzen 1993b.)

not significantly distinguish one gain from another.[2] The only possible distinction that can be noted in Figure 15 is that the response for strong negative feedback (gain < 1) tends to peak a year following eruption, while the response for positive feedback (gain > 1) tends to peak two years after

[2] The claim (Hansen et al 1992) that the prediction of cooling following Pinatubo forms a test of a model, ignores the fact that such a prediction fails to distinguish between models that predict $\Delta T_{2CO_2} = 5°C$ and those that predict $\Delta T_{2CO_2} = 0.25°C$.

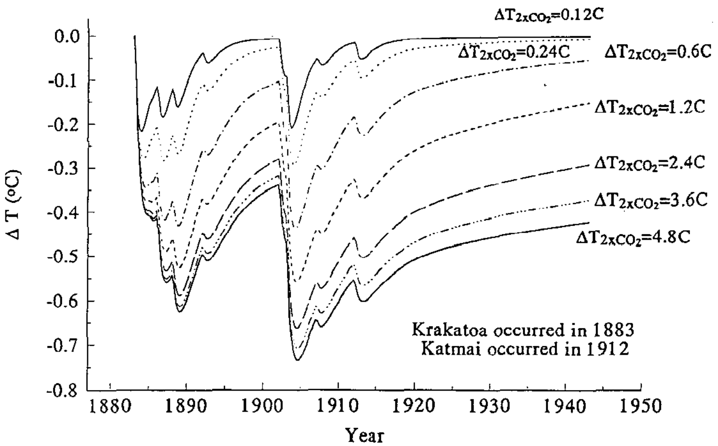

Figure 16 Response to series of volcanoes between Krakatoa in 1883 and Katmai in 1912. (From Lindzen 1993b.)

eruption. After 5–10 years, models with high sensitivity retain much more of the volcanic cooling than do models with low sensitivity. However, for individual volcanoes, the amount in either case is smaller than normal interannual variability. The situation is potentially better if one has a sequence of relatively closely spaced major volcanic eruptions, as was the case between Krakatoa in 1883 and Katmai in 1912 (Oliver 1976). Here, one might expect that for high sensitivity, the response of each volcano will add to the tail of the preceding volcano leading to a pronounced cooling trend. For low sensitivity, however, each volcano will produce an almost independent blip. The situation is illustrated in Figure 16. We see that high sensitivity leads to a pronounced cooling trend from 1883 to 1912 followed by persistent low temperatures (there were no further major volcanoes for about 50 years). Looking at Figure 1, we see only cooling blips corresponding to low sensitivity followed by pronounced warming between 1919 and 1940. It could of course be argued that the warming trend had begun earlier and had been cancelled by the volcanic cooling. Recall, however, that high sensitivity is associated with long delay so that the posited warming trend would have required forcing far greater than could have been accounted for by increasing CO_2 or any other known source.

6. REMARKS ON INCREASING CO_2

CO_2 in the atmosphere has increased from about 280 ppmv in 1800 to about 355 ppmv today. This represents an increase in excess of 25%. In addition, other minor greenhouse gases have also increased leading to an effective increase of about 50% in greenhouse gases. There is concern that continuing increases will lead to significant global warming. The focus of the present review has been on the basic physics of climate, and implications for climatic response to specified effective CO_2 increases. However, a few words are in order on how these increases have been specified. The reader may have noticed that the common measure of climate sensitivity is the equilibrium response to a doubling of CO_2, while the IPCC projections are based on the transient response to a quadrupling of effective CO_2 by 2100. As was noted in Section 5, high sensitivity is inevitably accompanied by long ocean delays. We have also seen that if the delays were shorter, then the record of the past century would have been inconsistent with high sensitivities. Given the need for long delays, the response to a mere doubling of CO_2 by 2100 would have been far smaller than the equilibrium responses; quadrupling was necessary for the transient response to be comparable by 2100 to the equilibrium response for a doubling of CO_2. Such a scenario was labeled by Houghton et al (1990) as the "business as usual scenario." However, it was, in fact a scenario designed to double effective CO_2 by 2030 and quadruple it later in the century. In order to arrive at such a scenario, it was necessary to project substantial increases in population, higher standards of living in the currently less developed world, increased reliance on coal, restrictions on nuclear power, etc. Recognition of the vast uncertainty of all projections over such long periods led to the presentation of a broad range of possibilities in Houghton et al (1992). It became clear that the main determinant of emissions would be population and economic growth in the currently less developed countries, and that emission controls in the currently developed countries was of relatively small long-term importance. In addition to socio-economic uncertainties, there are significant geochemical uncertainties in translating emissions into atmospheric CO_2.

The complexity and uncertainty of the chemistry are substantial (Heimann 1992). Here, we simply wish to consider some broad aspects of the problem. Measurements of ice cores suggest that preindustrial levels of CO_2 were approximately 280 ppmv for at least hundreds of years. This is suggestive of an equilibrating process. The simplest representation of such a process is

$$\frac{dn_{CO_2}}{dt} = a(\bar{n}_{CO_2} - n_{CO_2}) + S_{CO_2}, \tag{5}$$

where n refers to density, the overbar to an equilibrium value, a^{-1} to an equilibration time, and S to a source. Equation 5 is of course largely schematic, but it provides a framework for discussion. Clearly, an S that remains constant, will simply lead eventually to a new equilibrium with none of the emissions contributing to additional atmospheric CO_2. However, from 1800 to 1973, CO_2 sources are believed to have been exponentially increasing with an e-folding time, τ, of about 45 years. Such a system cannot be in equilibrium; the degree of disequilibrium depends on the relative magnitude of τ and a^{-1}. If a^{-1} is much shorter than τ then the system will always be near equilibrium and very little of the increased emission will appear in the atmosphere. If, on the other hand, a^{-1} is much longer than τ, then most of S will appear in the atmosphere. In point of fact, about half the CO_2 provided by S has remained in the atmosphere, suggesting that a^{-1} and τ are comparable. One remarkable aspect of all the IPCC scenarios is that they all call for τ to be much larger than 45 years during the next century, and at the same time rather inconsistently have the fraction of CO_2 remaining in the atmosphere increasing as well. It would appear that one is exaggerating the atmospheric consequences of the uncertain emissions.

Of course, the questions posed in this review concerning how climate behaves are of immense importance regardless of the specific behavior of CO_2. However, there can be little question that without increasing minor greenhouse gases the political import of the question diminishes significantly.

7. SUMMARY AND REMARKS

This review has stressed the basic questions in climate that transcend the specific concern for the role of minor greenhouse gases: Namely, what determines the mean temperature of the Earth and its distribution—specifically the distribution with latitude? Major climate changes in the past have been characterized by large changes in the equator-to-pole temperature difference (ranging from about 19°C to 60°C; the present annually averaged value is about 40°C), and relatively constant equatorial temperatures (within about 2°C of the present). Changes in the mean temperature of the Earth appear to have been a by-product of these changes rather than a cause. The cause(s) for the near constancy of equatorial temperatures constitutes an additional major question. The stability of equatorial temperatures is particularly remarkable when one considers that the changes in heat flux out of the tropics implicit in the changes in equator-to-pole temperature difference are likely to have been far larger than any proposed external radiative forcing. Indeed, given the stability of the tropics, it is hard to see how gross external radiative forcing, which does not affect meridional heat fluxes in any evident way, can substantially alter the mean temperature of the Earth.

Indeed, despite an increase of effective CO_2 of about 50% over the past century, the data do not display any change that can be distinguished from normal climate variability. If the ocean delay is on the order of 160 years, then the observed warming over the past century of 0.45°C ± 0.15°C is "broadly" consistent with any ΔT_{2CO_2} between 0 and 5°C. However, if the delay times should prove shorter, then the larger values become progressively incompatible with the data. As we have noted, such data as one has from the sequence of volcanoes between Krakatoa (1883) and Katmai (1912) are suggestive of small delays and negative feedbacks.

It has been commonly suggested that it will be decades before anthropogenic warming will be identified in the data, and that it may take a similar length of time before large-scale computer simulations of climate become dependable. This may be true at some level. However, a distinctly more optimistic view is appropriate at another level. For example, our present fears of large warming are based on specific aspects of current models: most notably their treatment of upper level water vapor. Improved understanding of the water vapor budget based on both theory and thoughtful observational analyses (both of which can be achieved in a much shorter time) should enable us to determine whether current predictions have any substantial foundation. To be sure, there may exist hitherto unknown processes that could still lead to warming, but unknown processes offer little policy guidance. We have also noted that ocean delay offers a direct measure of feedback. It should be possible to explore existing data concerning this matter. Finally, our increasing knowledge of the Earth's past climate provides a valuable test-bed for our quantitative understanding of climate. Such understanding is not limited to the output of large-scale simulations.

ACKNOWLEDGEMENTS

The preparation of this review was supported by NASA under grant NAGW-525, and by the NSF under grant ATM-914441.

Literature Cited

Arakawa, A., Schubert, W. H. 1974. Interaction of a cumulus cloud ensemble with the large scale environment, Part I. *J. Atmos. Sci.* 31: 674–701

Arking, A. 1993. Water vapor and lapse rate feedback: insight from one-dimensional climate model. *J. Clim.* In press

Balling, R. C. Jr. 1992. *The Heated Debate.* San Francisco: Pacific Res. Inst. 195 pp.

Barron, E. J. 1987. Eocene equator-to-pole surface ocean temperatures: A significant climate problem? *Paleoceanogr.* 2: 729–39

Berger, A. L. 1978. Long-term variations of

daily insolation and Quaternary climatic changes. *J. Atmos. Sci.* 35: 2362–67
Betts, A. K. 1990. Greenhouse warming and the tropical water vapor budget. *Bull. Am. Meteorol. Soc.* 71: 1465–67
Boer, G. J., Arpe, K., Blackburn, M., Déqué, M., Gates, W. L., et al. 1992. Some results from an intercomparison of the climates simulated by 14 atmospheric general circulation models. *J. Geophys. Res.* 97: 12,771–86
Bolton, D. 1980. The computation of equivalent potential temperature. *Mon. Weather Rev.* 108: 1046–53
Budyko, M. I., Izrael, Y. A. 1991. In *Anthropogenic Climate Change*, ed. M. I. Budyko, Y. A. Izrael, pp. 277–318. Tucson: Univ. Ariz. Press
Carissimo, B. C., Oort, A. H., Vonder Haar, T. H. 1985. Estimating the meridional energy transports in the atmosphere and ocean. *J. Phys. Oceanogr.* 15: 82–91
Cess, R. D., Potter, G. L., Blanchet, J. P., Boer, G. J., Del Genio, A. D., et al. 1990. Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models. *J. Geophys. Res.* 95: 16,601–15
Cess, R. D., Potter, G. L., Zhang, M.-H., Blanchet, J.-P., Chalita, S., et al. 1991. Interpretation of snow-climate feedback as produced by 17 general circulation models. *Science* 253: 888–92
Charney, J. G. 1947. The dynamics of long waves in a baroclinic westerly current. *J. Meteorol.* 4: 135–63
Charney, J. G., Stern, M. E. 1962. On the instability of internal baroclinic jets in a rotating atmosphere. *J. Atmos. Sci.* 19: 159–72
Crowley, T. J., North, G. R. 1991. *Paleoclimatology*. New York: Oxford Univ. Press. 339 pp.
Eady, E. T. 1949. Long waves and cyclone waves. *Tellus* 1: 33–52
Elliot, W. P., Gaffen, D. J. 1991. On the utility of radiosonde humidity archives for climate studies. *Bull. Am. Meteorol. Soc.* 72: 1507–20
Emanuel, K. A. 1991. A scheme for representing cumulus convection in large-scale models. *J. Atmos. Sci.* 48: 2313–35
Evans, D. L., Freeland, H. J. 1977. Variations in the earth's orbit: Pacemaker of the ice ages? *Science* 198: 528–30
Geleyn, J.-F., Girard, C., Louis, J.-F. 1982. A simple parameterization of moist convection for large-scale atmospheric models. *Beitr. Phys. Atmosph.* 55: 325–34
Goody, R. M., Yung, Y. L. 1989. *Atmospheric Radiation*. New York: Oxford Univ. Press. 519 pp.
Hansen, J., Lacis, A., Ruedy, R., Sato, M. 1992. Potential climate impact of Mount Pinatubo eruption. *Geophys. Res. Lett.* 19: 215–18
Hansen, J., Russell, G., Lacis, A., Fung, I., Rind, D. 1985. Climate response times: dependence on climate sensitivity and ocean mixing. *Science* 229: 857–59
Heimann, M. 1991. *Modelling the global carbon cycle*. Presented at First Demetra Meeting on Climate Variability and Global Change, Chianciano Therme, Italy, Oct. 28–Nov. 3, 1991. Proceedings of Meeting in preparation
Hou, A. Y. 1993. The influence of tropical heating displacements on the extratropical climate. *J. Atmos. Sci.* In press
Hou, A. Y., Lindzen, R. S. 1992. Intensification of the Hadley circulation due to concentrated heating. *J. Atmos. Sci.* 49: 1233–41
Houghton, J. T., Callander, B. A., Varney, S. K. 1992. Update to *Climate Change* 1992, *Suppl. Rep. IPCC Sci. Assessment*. Cambridge: Cambridge Univ. Press. 200 pp.
Houghton, J. T., Jenkins, G. J., Ephraums J. J., eds., 1990. *Climate Change. The IPCC Scientific Assessment*. Cambridge: Cambridge Univ. Press. 365 pp.
Imbrie, J., Imbrie , K. P. 1980. *Ice Ages, Solving the Mystery*. Hillside, NJ: Enslow. 213 pp.
Lindzen, R. S. 1988. Some remarks on cumulus parameterization. *Pageoph* 126: 123–35
Lindzen, R. S. 1990. *Dynamics in Atmospheric Physics*. New York: Cambridge Univ. Press. 310 pp.
Lindzen, R. S. 1993a. Baroclinic neutrality and the tropopause. *J. Atmos. Sci.* 50: 1148–51
Lindzen, R. S. 1993b. Constraining possibilities versus signal detection. In *Review of Climate Variability on the Decade to Century Time Scale*, National Research Council. To appear
Lindzen, R. S., Farrell, B. 1980. The role of polar regions in global climate, and the parameterization of global heat transport. *Mon. Weather Rev.* 108: 2064–79
Lindzen, R. S., Hou, A. Y. 1988. Hadley circulations for zonally averaged heating centered off the equator. *J. Atmos. Sci.* 45: 2416–27
Lindzen, R. S., Pan, W. 1993. A note on orbital control of equator-pole heat fluxes. *Clim. Dyn.* In press
Lorenz, E. N. 1967. *The Nature and Theory of the General Circulation of the Atmosphere*. Geneva: World Meteorol. Organ. 161 pp.
Milankovitch, M. 1930. *Mathematische Klimalehre und Astronomische Theorie der*

Klimaschwankungen. Berlin: Gebrüder Borntraeger 176 pp.

Mitchell, J. F. B., Senior, C. A., Ingram, W. J. 1989. CO_2 and climate: a missing feedback? *Nature* 341: 132–34

Möller, F., Manabe, S. 1961. Über das Strahlungsgleichgewicht der Atmosphäre. *Z. Meteorol.* 15: 3–8

Oliver, R. C. 1976. On the response of hemispheric mean temperature to stratospheric dust: an empirical approach. *J. Appl. Meteorol.* 15: 933–50

Oort, A. H. 1983. Global atmospheric circulation statistics, 1958–1973. *NOAA Prof. Pap.* 14. Rockville, Md.: NOAA, U.S. Dept. Commerce. 180 pp.

Oort, A. H. 1993. Observed humidity trends in the atmosphere. In *Proc. 17th Annu. Climate Dynamics Workshop*. To appear

Oort A. H., Rasmussen, E. M. 1970. On the annual variation of the monthly mean meridional circulation. *Mon. Weather Rev.* 98: 423–42

Peixoto, J. P., Oort, A. H. 1992. *Physics of Climate*. New York: Am. Inst. Phys. 520 pp.

Pocinki, L., 1955. Stability of a simple baroclinic flow with horizontal shear. *A. F. Cambridge Res. Center, Res. Pap. No.* 38. 78 pp.

Ramanathan. V., Collins, W. 1991. Thermodynamic regulation of ocean warming by cirrus clouds deduced from the 1987 El Niño. *Nature* 351: 27–32

Raval, A., Ramanathan, V. 1989. Observational determination of the greenhouse effect. *Nature* 342: 758–61

Rind, D., Chiou, E. W., Chu, W., Larsen, J., Oltmans, S., et al. 1991. Positive water vapor feedback in climate models confirmed by satellite data. *Nature* 349: 500–3

Sarachik, E. S. 1985. A simple theory for the vertical structure of the tropical atmosphere. *Pageoph* 123: 261–71

Smith, R. B. 1992. Deuterium in North Atlantic storm tops. *J. Atmos. Sci.* 49: 2041–57

Spencer, R. W., Christy, J. R. 1990. Precise monitoring of global temperature trends from satellites. *Science* 247: 1558–62

Stone, P. H. 1978. Baroclinic adjustment. *J. Atmos. Sci.* 35: 561–71

Stone, P. H., Risbey, J. 1990. On the limitations of general circulation models. *Geophys. Res. Lett.* 17: 2173–76

Sun, D.-Z., Lindzen, R. S. 1993a. Negative feedback of water vapor inferred from the mountain snowline record. *Ann. Geophys.* 11: 204–15

Sun, D.-Z., Lindzen, R. S. 1993b. On the distribution of tropical tropospheric water vapor. *J. Atmos. Sci.* 50: 1643–60

Sun, D.-Z., Lindzen, R. S. 1993c. An EPV view of the zonal mean distribution of temperature and wind in the extra-tropical troposphere. *J. Atmos. Sci.* Submitted

Trenberth, K. E., Christy, J. R., Hurrell, J. W. 1992. Monitoring global monthly mean surface temperatures. *J. Clim.* 12: 1405–23

Williams, E. R., Rutledge, S. A., Geotis, S. G., Renno, N., Rasmussen, E., et al. 1992. A radar and electrical study of tropical "hot towers." *J. Atmos. Sci.* 49: 1386–95

Winograd, I. J., Coplen, T. B., Landwehr, J. M., Riggs, A. C., Ludwig, K. R., et al. 1992. Continuous 500,000-year climate record from vein calcite in Devils Hole, Nevada. *Science* 258: 255–60

Annual Review of Fluid Mechanics
Volume 26, 1994

CONTENTS