Chapter 2

Climate Sensitivity to Increasing Greenhouse Gases

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INTRODUCTION

Climate changes occur on all time scales, as illustrated in Figure 2-1 by the trend of global mean surface air temperature in the past century, the past millennium, and the past 30,000 years. The range of global mean temperature in the past 30,000 years and indeed the past million years has been of the order of 5ºC. At the peak of the last glacial period, the Wisconsin ice age approximately 18,000 years ago, the mean temperature was 3-5ºC (5-9ºF) cooler than today. At the peak of the current interglacial, 5,000-8,000 years ago, the mean temperature is estimated to have been 0.5-1ºC warmer than today (Figure 2-1). In the previous (Eemian) interglacial, when sea level is thought to have been about 5m higher than today (Hollin, 1972), global mean temperature appears to have been of the order of 1ºC warmer than today.

Global mean temperature is a convenient parameter, but it must be recognized that much larger changes may occur on more localized scales. Decadel variations of global temperature in the past century, for example, are enhanced by about a factor of three at high latitudes (Hansen et al., 1983a). Also, the global cooling of 3-5ºC (5-9ºF) during the Wisconsin ice age included much larger regional changes, as evidenced by the ice sheet of 2 km (1.3 mi) mean thickness covering much of North America including the present sites of New York, Minneapolis, and Seattle.

The recorded climate variations include the response to external forcings (e.g., changes in the amount or global distribution of solar irradiance) and also internal climate fluctuations (e.g., changes in ocean dynamics driven by weather "noise"). Determination of the division of actual climate variations between these two categories is a fundamental task of climate investigations.

The mean temperature of the earth is determined primarily by the amount of energy absorbed from the sun, which must be balanced on average by thermal emission. The earth's surface temperature also depends on the atmosphere, which partially blankets the thermal radiation and thus requires the surface to be hotter in order for the thermal emission to balance the absorbed solar radiation. Today the mean temperature of the earth's surface is 288K, 33EC higher than it would be in the absence of this "greenhouse" blanketing by the atmosphere.

As the $C0_2$ content of the atmosphere increases, the atmosphere becomes more opaque at infrared wavelengths where $C0_2$ has absorption bands, thus raising the mean level of emission to space to higher altitudes. A simple radiative calculation shows that doubling atmospheric $C0_2$ would raise the mean level of emission to space, averaged over the thermal emission spectrum, by about 200m. (Cf. discussion in the section below on empirical evidence of climate sensitivity.) Since atmospheric temperature falls off with altitude by about 6EC/km, the planet would have to warm by about 1.2EC to restore equilibrium if the tropospheric temperature gradient and other factors remained unchanged. In general, other factors would not remain unchanged, and thus the actual temperature change at equilibrium would differ from the one in this simple calculation by some "feedback" factor, *f*,

$$) I_{eq} = f) I_{rad} \qquad (2.1)$$

where $) I_{eq}$ is the equilibrium change in global mean surface air temperature and $) I_{rad}$ is the change in surface temperature that would be required to restore radiative equilibrium if no feedbacks occurred.

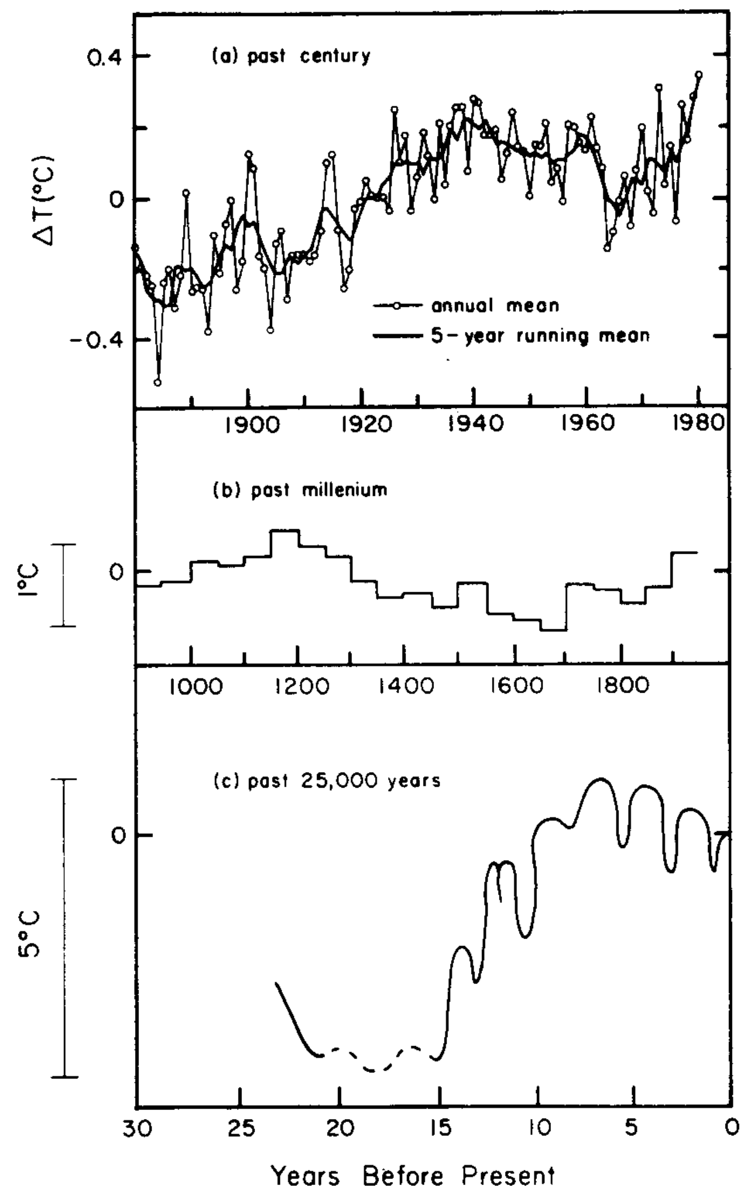

Figure 2-1. Global temperature trend for *(a)* past century, *(b)* millennium, and *(c)* 25,000 years. *(a)* is based on J. Hansen, D. Johnson, A. Lacis, S. Lebedeff, P. Lee, D. Rind, and G. Russell, 1981, "Climate Impact of Increasing Atmospheric Carbon Dioxide," Science **213:**957-966, updated through 1981. (b) is based on temperatures in central England, the tree limit in the White Mountains of California, and oxygen isotope measurements in the Greenland ice (W. Dansgaard of the Geophysical Isotope Laboratory, University of Copenhagen, pers. comm.), with the temperature scale set by the variations in the last 100 years. (c) is based on changes in tree lines, fluctuations of alpine and continental glaciers, and shifts in vegetation patterns recorded in pollen spectra (National Academy of Sciences, *1975. Understanding Climatic Change,* Washington, D.C.: National Academy Press), with the temperature scale set by the 3-5° cooling obtained in a 3-D climate model J.E. Hansen et al., 1983*b*, "Efficient Three Dimensional Global Models for Climate Studies: Models I and II,*" Monthly Weather Review* **111:**609662; J. Hansen et al., 1984; "Climate Sensitivity: Analysis of Feedback Mechanisms," in *Climate Processes and Climate Sensitivity,* J.E. Hansen and T. Takahashi, eds., Washington, D.C.: American Geophysical Union, pp. 130-163 with the boundary conditions for 18,000 years ago. Thus, the shapes of curves *(b)* and *(c)* are based on only Northern Hemisphere data.

The feedback factor *f* not only determines the magnitude of the eventual climate change for a given change in climate forcing but also the time required to approach the new equilibrium. The reason for this is the fact that the initial rate at which the ocean warms is determined by only the magnitude of the direct climate forcing, that is, the feedbacks only come into play as the warming occurs, and thus the ocean thermal response time increases with increasing *f* (Hansen et al., 1981, 1984). The physical processes expected to contribute to the feedback factor include the ability of the atmosphere to hold more water vapor (which is also a greenhouse gas) with increasing temperature and the change of snow and ice cover (and thus albedo) with changing temperature.

In this chapter we first discuss current climate model evidence for climate sensitivity, which suggests a range of 3±1.5EC for doubled $C0_2$, corresponding to a net feedback factor *f* - 2.5. We then summarize empirical evidence for climate sensitivity and feedback processes, which provide substantial support for the magnitude of climate effects computed by the models. Finally, we look at current trends of greenhouse gases and global temperature, which allow us to discuss the magnitude of warming expected in coming decades.

CLIMATE MODEL CALCULATIONS OF CLIMATE SENSITIVITY

Two National Academy of Sciences panels (Charney, 1979; Smagorinsky, 1982) estimated the equilibrium (global mean) climate sensitivity for doubled $C0_2$ to be 3±1.5EC. This conclusion was based on consideration of the primary mechanisms believed to contribute to global climate sensitivity, including the study of results from two 3-D global climate models: that of Manabe and Stouffer (1980), which yields a sensitivity near 2EC, and that of Hansen et al.(1983*b*, 1984), which yields a sensitivity near 4EC.

In this section we illustrate the temperature change produced in the latter 3-D climate model when $C0_2$ is doubled and analyze the physical mechanisms contributing to this sensitivity. This provides a basis for discussing the uncertainties in the computed climatic sensitivity due to approximations in representing these processes. The temperature change computed for doubled $C0_2$ is shown in Figure 2-2 in three ways: *(a)* the annual mean surface air temperature as a function of latitude and longitude; *(b)* the zonal mean surface air temperature as a function of latitude and month; and *(c)* the annual and zonal mean temperature as a function of altitude and latitude.

The surface air warming is enhanced at high latitudes (Figure 2-2*a*) partly because of confinement of the greenhouse warming to lower layers as a consequence of the atmospheric stability at high latitudes and partly because of the ice/snow albedo feedback at high latitudes. The enhanced warming in the African and Australian deserts can be traced to the stability of the atmosphere above these regions and the relative lack of evaporative cooling. The maximum warming near West Antarctica is associated with the largest reduction in sea ice cover there. Many aspects of the geographical distribution of the warming for doubled $C0_2$ are clearly related to changes in prevailing wind patterns. However, the detailed geographical patterns of the computed climate changes should not be viewed as a reliable prediction for a doubled $C0_2$ world, because current climate models still poorly represent many parts of the climate system. For example, changes in horizontal heat transport by the oceans, which will undoubtedly influence regional climate patterns, are not included in the simulations for doubled $C0_2$.

The strong seasonal variation of the computed warming at high latitudes (Figure 2-2*b*) is due to the seasonal change of atmospheric stability and the influence of melting sea ice in the summer, which limits the ocean temperature rise. At low latitudes, the temperature rise is greatest in the upper troposphere (Figure 2-2*c*), because the added heating at the surface primarily causes increased evaporation and moist convection, with resultant deposition of latent heat and water vapor at high levels. These processes are discussed further below.

Figure 2-2. Warming of air temperature due to doubled $C0_2$ in the 3-D global climate model of Hansen et al. *(a)* shows the geographical

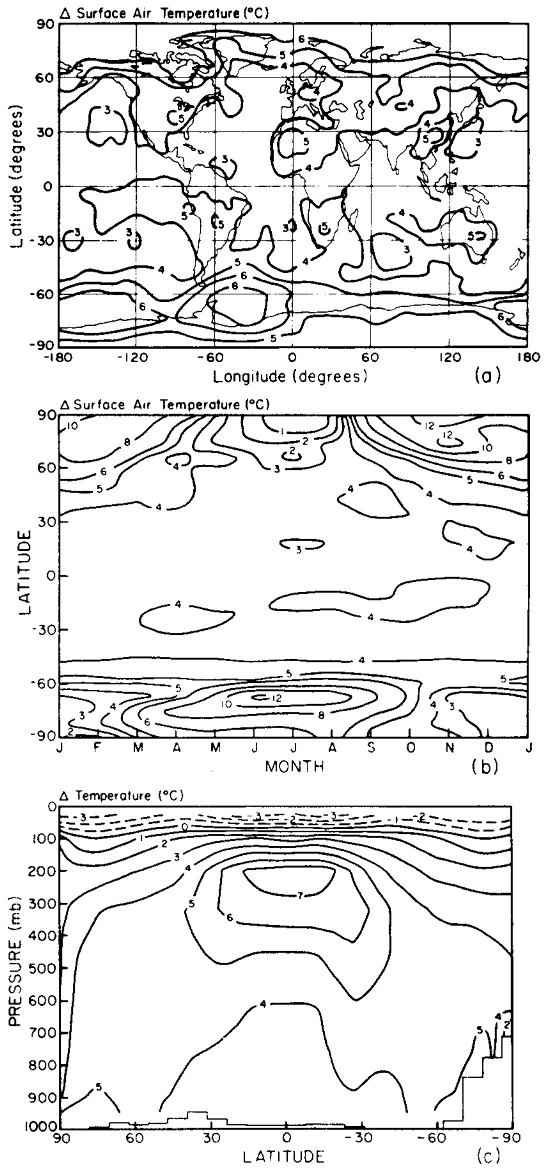

Figure 2-2. Warming of air temperature due to doubled CO_2 in the 3-D global climate model of Hansen et al. (*a*) shows the geographical distribution of annual mean surface air warming; (*b*) shows the seasonal variation of the surface air warming averaged over longitude; and (*c*) shows the altitude distribution of the warming averaged over season and longitude. (After J. Hansen, A. Lacis, D. Rind, G. Russell, P. Stone, I. Fung, R. Ruedy, and J. Lerner, 1984. "Climate Sensitivity: Analysis of Feedback Mechanisms." in *Climate Processes and Climate Sensitivity,* J. E. Hansen and T. Takahashi, eds., Washington, D.C.: American Geophysical Union, pp. 130–163.)

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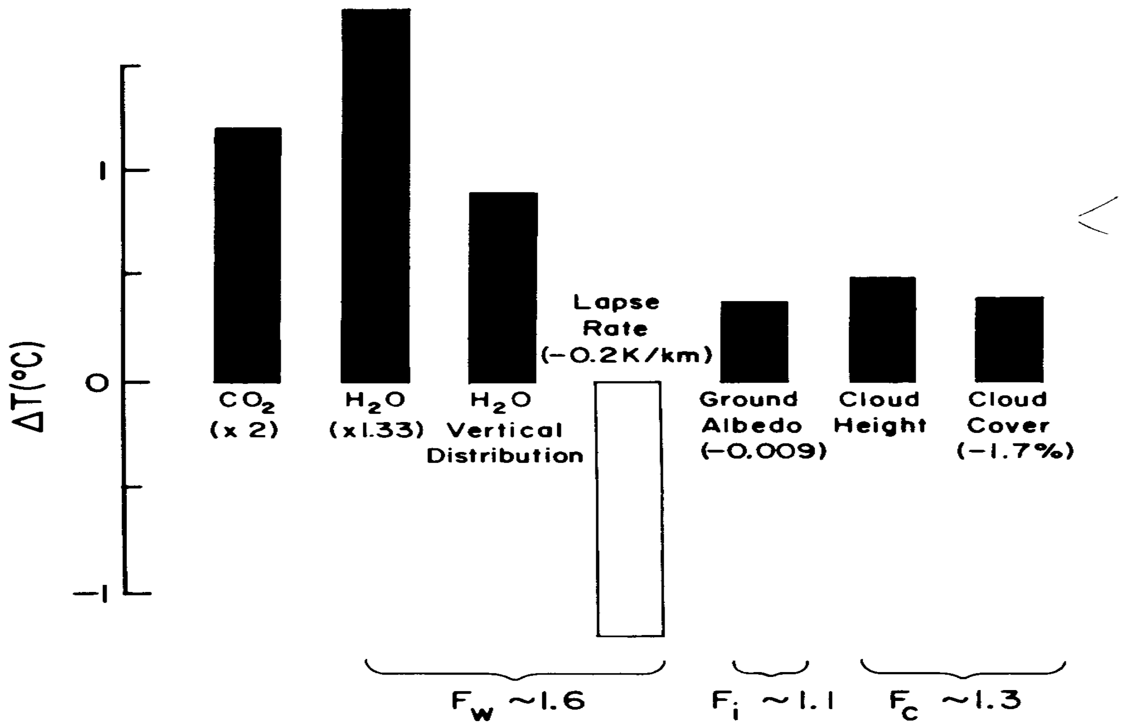

Figure 2-3. Contributions to the global mean temperature rise in the doubled CO_2 experiment as estimated by inserting changes obtained in the 3-D experiment into a 1-D radiative/convective climate model.

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The processes in this 3-D climate model that are responsible for the 4EC temperature rise for doubled $C0_2$ can be investigated with the help of a 1-D model of Lacis et al. (1981), inserting into it one-by-one all the radiatively significant global mean changes that were observed to occur in the 3-D experiments. This analysis procedure and its limitations are discussed in greater detail elsewhere (Hansen et al., 1984). The contributions found for the different changes that occurred in the 3-D model are illustrated in Figure 2-3.

Water vapor increase provides the dominant feedback, with most of the effect given by the increase in mean water vapor amount. Additional positive feedback occurs because the water vapor distribution is weighted more to higher altitudes for the doubled $C0_2$ case. However, the change in lapse rate, mainly due to the added H_20, almost cancels the effect of the change in the water vapor vertical profile. Since the amount of water the atmosphere holds is largely dependent on the mean temperature, it is expected that the latter two effects would approximately cancel. Thus, it seems unlikely that the net water vapor feedback factor can be greatly in error, even though the water vapor distribution and lapse rate depend on the moist convection process, which is difficult to model realistically.

Ground albedo decrease also provides a substantial feedback (Figure 2-3). The ground albedo change (Figure 2-4*a*) is largely due to reduced sea ice. Shielding of the ground by clouds and the atmosphere (Figure 2-4*b*) makes this feedback several times smaller than it would be in the absence of the atmosphere. However, it is a significant positive feedback, and for this model it is at least as large in the Southern Hemisphere as in the Northern Hemisphere. The 1-D RC model does not provide a complete analysis of the sea ice feedback, for example, of the effect of sea ice in insulating the ocean and thus reducing radiation to space. However, from the geographic pattern of the temperature increase (Figure 2-2), and the coincidence of warming maxima with reduced sea ice, it is clear that the sea ice provides a positive feedback.

Cloud changes (Figure 2-5) also provide a significant positive feedback for doubled $C0_2$ in this model, as a result of a small increase (about 10mb) in mean cloud height and a 1.7 percent decrease in cloud cover. Present understanding of cloud processes does not permit confirmation or contradiction of the realism of these changes. The increase of mean cloud height is plausible: it falls between the common assumptions of fixed cloud height and fixed cloud temperature. The change in cloud cover reflects a reduction of low and middle level clouds, associated with a drying of these layers due to an increase of penetrating cumulus convection. Improved assessment of the cloud contribution will depend primarily on the development of increasingly realistic representations of cloud formation processes in global climate models, as verified by an accurate global cloud climatology.

The processes providing the major feedbacks in this climate model are thus atmospheric water vapor, clouds, and the surface albedo. Considering the earth from a planetary perspective, it seems likely that these are the principal feedback processes for the earth on a time scale of decades. The albedo of the planet for solar radiation is primarily determined by the clouds and the surface, with the main variable component of the latter being the ice/snow cover. The thermal emission of the planet is primarily determined by the atmospheric water vapor and clouds. Thus, those processes principally responsible for the earth's radiation balance and temperature are included in the model and are responsible for the significant feedbacks in the model.

There is substantial uncertainty in the quantitative value of these feedbacks. However, the most important feedback, due to water vapor, seems certain to be greater than one and is unlikely to be less than approximately 1.5. The ice/snow albedo feedback seems certain to be greater than one. The cloud feedback could be greater or less than one. Our model suggests that it is a significant positive (f>1) feedback, but much more work is needed.

These feedback factors suggest some sources for the difference between our climate model sensitivity and that of Manabe and Stouffer (1980). They use fixed clouds (altitude and cloud cover) and thus have f_{cloud}/1. Also, their control run has less sea ice than our model, so that their feedback factor for that process should be between one and the value for our model. Therefore, it is likely that their primary feedback is $f_{water\ vapor}$, and it is not surprising that their sensitivity is approximately 2EC for doubled $C0_2$.

Although the cloud and sea ice feedbacks appear to "account" for most of the difference in sensitivity between our model and that of Manabe and Stouffer, we point out that there is another major difference between the models. This difference relates to the atmosphere and ocean transports of energy, whose feedbacks do not show up as identified components in an energy balance analysis such as in Figure 2-3. Our model includes a specified horizontal transport of heat by the ocean, which is identical in the control and experiment runs; thus there is no ocean tranport feedback in our model.

Manabe and Stouffer do not explicitly allow feedback on ocean transport either, because the ocean transport is zero in both experiment and control runs. However, in their model, increased poleward transport of energy in the atmosphere apparently replaces poleward transport of heat in the ocean, since their high latitude regions are at least as warm as in our model (and observations). This surrogate oceanic transport (in the atmosphere) may provide a negative feedback; the decrease in latitudinal temperature gradient accompanying a warmer atmosphere generally tends to decrease atmospheric transports, thus providing a negative feedback (Stone, 1984). Thus, while our ocean transport has no feedback effect, being identical in experiment and control runs, Manabe and Stouffer's surrogate transport probably has a negative feedback; indeed, Manabe and Wetherald (1975, 1980) explicitly show a negative feedback poleward of mid-latitudes for doubled $C0_2$ runs with idealized topography. The contribution of this feedback could be quantified by running the same model with and without fixed ocean transport.

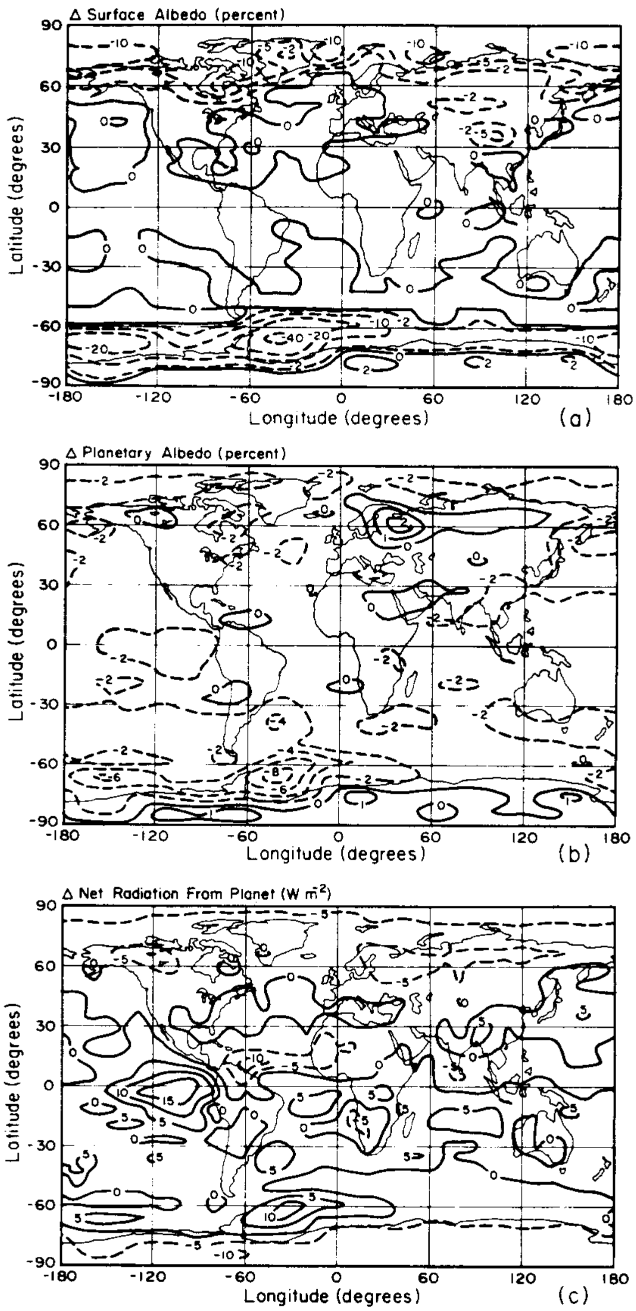

Figure 2-4. Geographical distribution of the annual mean changes of surface albedo, planetary albedo, and net radiation from the planet due to doubled CO_2 in the 3-D global climate model of Hansen et al. (After J. Hansen, A. Lacis, D. Rind, G. Russell, P. Stone, I. Fung, R. Ruedy, and J. Lerner, 1984. "Climate Sensitivity: Analysis of Feedback Mechanisms." in *Climate Processes and Climate Sensitivity,* J.E. Hansen and T. Takahaski, eds., Washington, D. C.: American Geophysical Union, pp. 130–163.)

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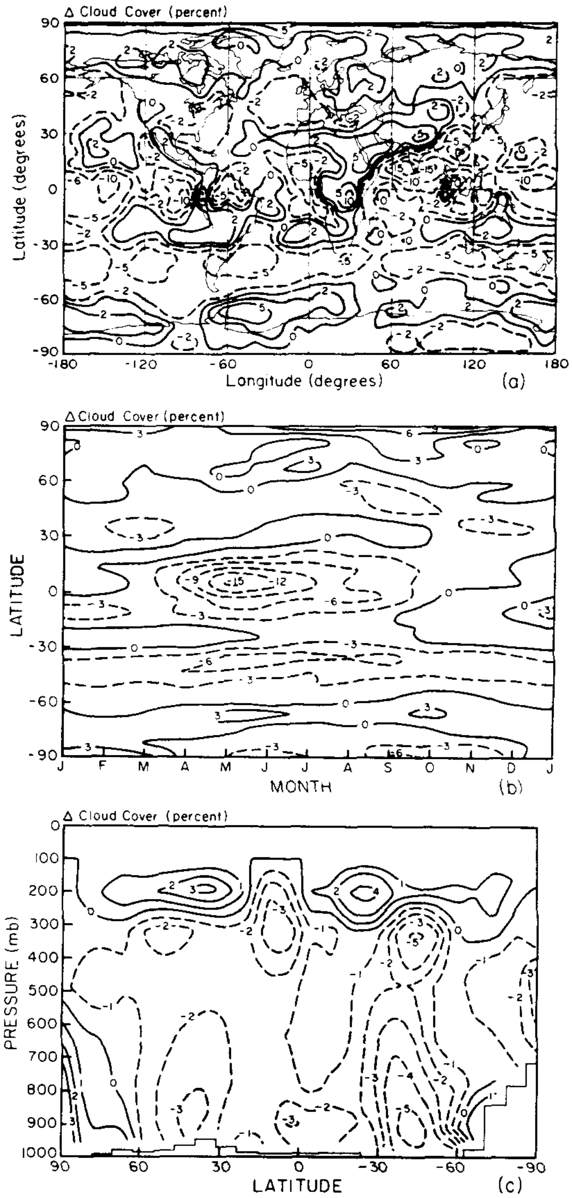

Figure 2-5. Cloud cover changes due to doubled CO_2 in the 3-D global climate model of Hansen et al. (*a*) show the geographical distribution of annual mean cloud cover changes; (*b*) shows the seasonal variation of cloud cover changes averaged over longitude; and (*c*) shows the altitude distribution of cloud cover changes, averaged over season and longitude. (After J. Hansen, A. Lacis, D. Rind, G. Russell, P. Stone, I. Fung, R. Ruedy, and J. Lerner, 1984. "Climate Sensitivity: Analysis of Feedback Mechanisms." in *Climate Processes and Climate Sensitivity*, J. E. Hansen and T. Takahashi, eds., Washington, D. C.: American Geophysical Union, pp. 130–163.)

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In summary, available global climate models all suggest an equilibrium global climate sensitivity in the range of 2-4EC for doubled $C0_2$. This range is consistent with that estimated by the National Academy of Sciences, 1.5-4.5EC, which attempted to allow for uncertainties not accounted for in existing models. It is certainly conceivable that the true climate sensitivity is outside this range. However, a sensitivity smaller than 1.2EC would require the hypothesis of a net negative feedback. Such a hypothesis, though it can not be ruled out a priori, is faced with the difficult task of finding a negative feedback strong enough to overcome the dominant feedback mechanism that has been identified, that is, the ability of the atmosphere to hold more water vapor at higher temperatures, which is strongly positive ($f_{\text{water vapor}}$ - 1.6).
Improvement of the ability of global climate models to realistically
simulate climate change will require better understanding of key physical processes such as moist convection, large-scale cloud formation and ocean circulation, including its response to a warming of the ocean mixed layer. Better understanding of these processes, in turn, depends on appropriate observations from both global-scale and small-scale studies.

EMPIRICAL EVIDENCE OF CLIMATE SENSITIVITY

Planetary Data

A valuable test of the magnitude of the greenhouse effect can be obtained by examining the ensemble of experiments provided by the conditions on several different planets. The terrestrial planets Mars, Earth, and Venus provide a particularly appropriate set because they have a broad range of abundances of greenhouse gases. We first summarize the nature of the greenhouse effect and then compare its magnitude on these planets.

The temperature of a planet is determined by the requirement that, averaged over time, the infrared emissions to space balance the absorbed solar radiation. The effective radiating temperature of the planet is obtained by equating the thermal emission to that of a blackbody, thus

$$BR^2(1\text{-}A)S_o=4BR^2FI_e^4 \qquad (2.2)$$

or

$$I_e=[S_o(1\text{-}A)/4F]^{1/4} \qquad (2.3)$$

where R is the planet's radius, A its albedo, S_o the flux of solar radiation and F the Stefan-Boltzmann constant.

The difference between the surface temperature and effective temperature of a planet, I_s - I_e, is the greenhouse effect due to the gaseous atmosphere and clouds that cause the mean radiating level to space to be at some altitude above the surface.

A quantitative estimate of the greenhouse effect can be obtained under the assumption that only radiation contributes significantly to vertical energy transfer. The Eddington approximate solution of the radiative transfer equation is

$$I_s=I_e(1+\tfrac{3}{4}J_e)^{1/4} \qquad (2.4)$$

with the effective infrared optical thickness J, obtained from

$$e^{-Je}=\text{I}e^{-Je}B(I_e)d{<}\text{I}B(I_e)d{<} \qquad (2.5)$$

where optical thickness is a dimensionless number such that the fraction exp ($-I_p$) of radiation impinging perpendicularly is transmitted without interaction, B is the Planck blackbody function and the integrations are over all frequencies <. However, if the atmosphere is sufficiently opaque in the infrared, the purely radiative vertical temperature gradient, dI/dh, may be so large as to be unstable, thus giving rise to atmospheric motions that contribute to vertical transport of heat. In that case a better estimate of the greenhouse warming can be obtained from

$$\mathrm{I}_s - I_e + {}'H \qquad (2.6)$$

where H is the altitude of the mean radiation level and ' is the measured or estimated mean temperature gradient (lapse rate) in the region between the surface and the mean radiating level.[1]

The quantitative theory for the greenhouse mechanism is tested by comparing Mars, Earth, and Venus in Table 2-1. In comparison to Earth, Mars has a small amount of infrared absorbing gases in its atmosphere, while Venus has a dense opaque atmosphere composed mainly Of $C0_2$. The relatively transparent atmosphere of Mars should cause a greenhouse effect of only a few degrees, as indicated by equation (4). On Earth the lowest few kilometers of the atmosphere are too opaque for pure radiative transfer of heat with a stable lapse rate as a result of the radiative opacity of clouds and the large amount of water vapor in the lower atmosphere. The mean lapse rate in the convectively unstable region is ' - 5.5EC km^{-1}, which is less than the dry adiabatic value (-10EC km^{-1}) because of the effects of latent heat release by condensation as moist air rises and cools and because the atmospheric motions that transport heat vertically include not only local convection but also large-scale atmospheric dynamics. The mean radiating level is in the mid-troposphere,[2] at altitude H - 6 km. The atmosphere of Venus is opaque to infrared radiation for most of the region between the surface and the cloud tops as a result Of $C0_2$, H_20, and aerosol absorption. The lapse rate is essentially the dry adiabatic value (-7EC km^{-1}) because of the absence of large latent heat effects and because the large-scale dynamics are unable to produce lapse rates that are appreciably subadiabatic (Stone, 1975). The cloud tops radiating to space are at altitude H -70 km.

The observed surface temperatures of Mars, Earth, and Venus are all consistent with the calculated greenhouse warming (Table 2-1). The planets thus confirm the existence and order of magnitude of the greenhouse effect. These checks of the greenhouse mechanism refer to cases that have had sufficient time to reach thermal equilibrium with space.

Paleoclimate Data

Substantial information exists on the nature of large climate changes that have occurred in the past on Earth. Knowledge is available, for example, of the areas covered by past ice sheets, as evidence by their scouring of rocks; of the areas of sea ice cover, as evidenced by ocean bottom sediments formed by detritus deposited by melting ice; and of oceanic and atmospheric temperatures, as evidenced by the isotopic composition and geographic distribution of organisms that grew at those times.

Ideally, we would like to have an accurate knowledge of all climate forcings on paleoclimate time scales and of the climate response, which would give us a direct empirical calibration of climate sensitivity. But we do not know all the forcings and the one that is precisely known, variations of solar irradiance due to Earth orbital fluctuations, is a subtle forcing arising from changes in the seasonal and geographic distribution of radiation rather than a net global change of total incoming flux. Thus, sophisticated models and improved climate data will be needed for the paleoclimate data to yield a direct calibration of climate sensitivity.

However, we can examine the contribution of certain feedback processes to paleoclimate temperature change and thus obtain a valuable measure of the contribution of these feedbacks to climate sensitivity. In particular, the CLIMAP (climate mapping) project (Denton and Hughes,1981) has determined detailed boundary conditions (see surface temperature, ice sheet coverage and altitude, land boundaries, and sea ice cover) for the Wisconsin ice age (approximately 18,000 years ago).

When the CLIMAP boundary conditions are inserted in a general circulation model (Hansen et al., 1984) they yield a global mean surface air temperature 4EC colder than either today's observed temperature or the temperature produced by the same model with today's boundary conditions. The uncertainty in this cooling is on the order of 1EC and is mainly due to uncertainty in the reconstructed ocean surface temperature, since the model results are fixed closely by the specified boundary conditions.

We can examine sea ice, land ice, and vegetation feedbacks individually by replacing CLIMAP 18K boundary specifications by today's conditions and computing the change in flux at the top of the atmosphere. Since the model sensitivity for a flux imbalance at the top of the atmosphere is known, it is possible to estimate the feedback factor for each of these processes in this way. We present elsewhere (Hansen et al.,

Table 2-1. Greenhouse Effect on Terrestrial Planets

	Observed or Estimated						T_S(°K)		
							Computed		
Planet	S_0($W\,m^{-2}$)	A	τ	Γ(°C km^{-1})	H(km)	T_e(°)	Eq.(1)	Eq.(2)	Observed
Mars	589	0.15	~0.1	5	1	217	221	222	~220
Earth	1367	0.30	~1	5.5	6	255	293	288	288
Venus	2613	0.75	≳100	7	70	232	≳685	720	~700

S_0 = solar irradiance
A = planetary albedo
τ = atmospheric infrared opacity
Γ = atmospheric mean lapse rate
H = mean altitude of emission to space
T_e = effective temperature = $[S_0(1 - A)/4\sigma]^{1/4}$
(expected planetary temperature in the absence of a greenhouse effect)
T_s = surface temperature

Radiative equilibrium

$$T_s = T_e\,(1 + \tfrac{3}{4}\tau)^{1/4} \tag{1}$$

Convective equilibrium

$$T_s = T_e + \Gamma H \tag{2}$$

1984) a quantitative evaluation of each of these feedback processes based on the CLIMAP boundary conditions.

The sea ice feedback implied by the changes in sea ice coverage during the last ice age is found to be a substantial positive feedback. This conclusion is consistent with the result obtained by the 3-D climate model in the doubled $C0_2$ experiment described above. This feedback operates on time scales comparable to or longer than the ocean surface temperature response time, and thus it should be included in estimates for the effects of increased $C0_2$ /trace gas abundances on decade-to-century climate change.

The CLIMAP data do not include cloud cover or atmospheric water vapor; therefore, these mechanisms are not tested by the paleoclimate data. However, the total climate sensitivity can be tested empirically on the basis of observed global temperature variations during the past century, as discussed below.

Recent Global Temperature Trends

Surface air temperature observations have been sufficiently widespread to define the global temperature trend for most of the past century (Hansen et al., 1981). For the same time period, there are substantial observational data on some of the principal global climate-forcing mechanisms. Thus, it is possible to examine the observed trend for the presence of any response to observed forcings, and if it is found, to use this as one calibration of climate sensitivity.

An additional variable is introduced in studying the past century, because for such a short period it can not be assumed that the planet is in thermal equilibrium with space, that is, it is necessary to consider the transient response to variable climate forcing. The response time of surface air temperature to changed climate forcing is dependent on the ocean surface thermal response time, because of the close thermal coupling between the surface air and ocean mixed layer. Although the thermal relaxation time of the mixed layer alone is only several years, exchange of water between the mixed layer and deeper layers may delay the full surface response by decades or even centuries. Thus, the rate of this exchange is the additional variable.

Atmospheric $C0_2$ has been accurately monitored since 1958, during which time its growth (from 315 ppm in 1958 to 340 ppm today) has been about half of the $C0_2$ release by fossil fuel burning. If this pattern prevailed earlier, the 1880 $C0_2$ abundance was about 290 ppm. Possible effects of deforestation make the 1880 abundance uncertain by perhaps 10-20 pm, but this uncertainty does not greatly influence the analysis of effects on global temperature. The other climate forcing known to be significant during the past century, atmospheric aerosols produced by volcanic emissions, has been reasonably well-defined during this century on the basis of atmospheric transparency measurements. These measurements directly yield the climatically important quantity, visible aerosol optical depth; together with knowledge of intra- and inter-hemispheric mixing times, a useful definition of the volcanic aerosol climate forcing is obtained. Other possible global climate forcings include growth of trace gases, whose effect has become comparable to $C0_2$ in the past one or two decades but was probably small compared to $C0_2$ growth earlier (Lacis et al., 1981). Fluctuations in solar output may account for part of the observed climate variability, but adequate measurements are not available for the past century.

The observed global temperature trend for the past century is compared to climate model calculations for two different choices of global climate forcings in Figure 2-6. The observed temperature increase of the past century is matched by the model with $C0_2$ + volcanoes forcing with an equilibrium climate sensitivity of 2.8EC for doubled $C0_2$ and an exchange rate 6= 1.2 cm^2 s^{-1}. This value for 6, the effective vertical diffusion coefficient in the thermocline beneath the 100 m ocean mixed layer, is in the range suggested by current knowledge of oceanic mixing processes.

However, if the exchange rate is somewhat larger or smaller, a good fit to the observed temperature trend can still be obtained with a different value for the climate sensitivity. We find that with an exchange rate between the ocean mixed layer and thermocline based on passive tracers (6-1-2 cm^2s^{-1}), a climate sensitivity of 2-5EC is needed to provide best fit to the observed global temperature trend.

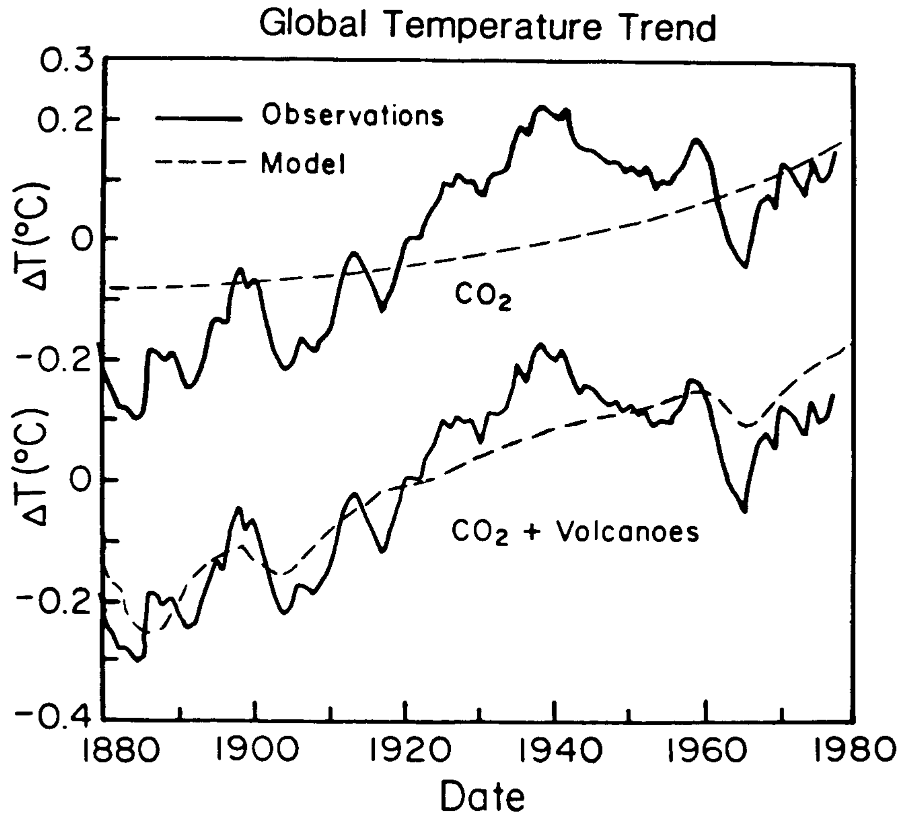

Figure 2-6. Global temperature trend computed with a climate model with sensitivity 2.8°C for doubled CO_2 and an exchange rate $\kappa = 1.2\ cm^2 s^{-1}$ between a 100 m mixed layer ocean and the thermocline. (After J. Hansen, D. Johnson, A. Lacis, S. Lebedeff, P. Lee, D. Rind, and G. Russell, 1981, "Climate Impact of Increasing Atmospheric Carbon Dioxide," *Science*, **213:**957-966.)

Figure 2-6. Global temperature trend computed with a climate model with sensitivity 2.8EC for doubled CO_2 and an exchange rate 6=1.2cm^{2}s^{-1} between a 100m mixed layer ocean and the thermocline. (After J. Hansen, D. Johnson, A. Lacis, S. Lebedeff, P. Lee, D. Rind, and G. Russell, 1981, "Climate Impact of Increasing Atmospheric Carbon Dioxide," *Science*, **213:**957-966.)

We conclude that the global temperature trend of the past century is generally consistent with the climate sensitivity estimated by the National Academy of Sciences committees (Charney, 1979; Smagorinsky, 1982): 1.5-4.5EC for doubled $C0_2$.

CURRENT TRENDS OF GREENHOUSE GASES AND GLOBAL TEMPERATURE

The average rate of global warming during the past century, 0.04EC per decade, can be expected to increase in the immediate future. This is because the absolute rate of the $C0_2$ increase (say in ppm/decade) is presently at its highest level and because other trace gases are now increasing at rates that significantly enhance the $C0_2$ warming.

The $C0_2$ increase in the decade of the 1970s was about 12 ppm, which was of the order of 25 percent of the total increase Of $C0_2$ for the period 1880-1980. In addition, several other trace gases increased in the 1970s by an amount sufficient to cause a greenhouse warming 50-100 percent as large as that due to $C0_2$ (Lacis et al., 1981 and Figure 2-7). The rates of change, of CH_4 and N_20 were not precisely measured for this period, but Figure 2-7 shows that even with conservative estimates for their growth rates (0.9 percent/yr for CH_4 and 0.2 percent/yr for N_20), the other trace gases yield a warming 70 percent as great as for $C0_2$ during the 1970s.

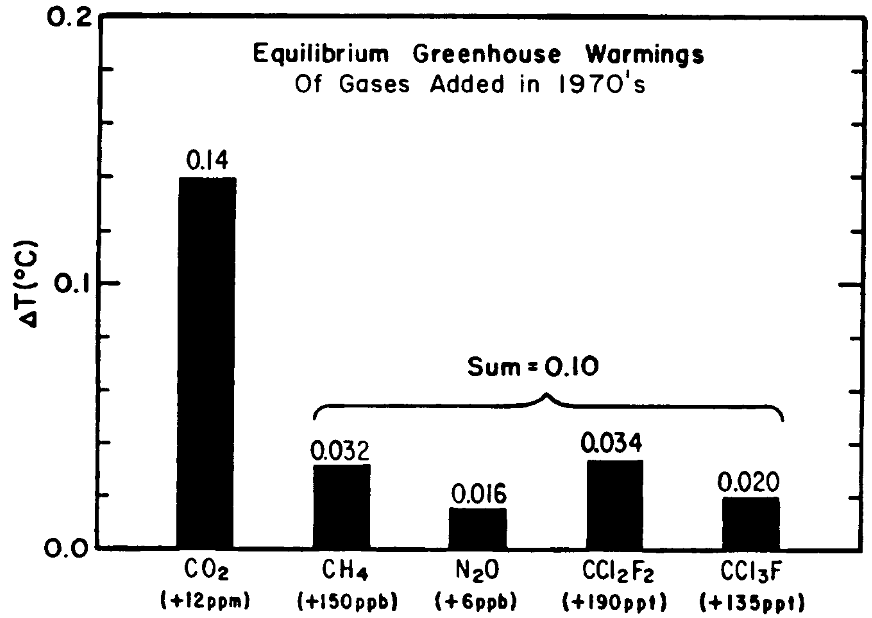

Figure 2-7. Equilibrium greenhouse warmings for estimated 1970–1980 abundance increases of several trace gases, based on climate model with sensitivity ~3°C for doubled CO_2. (After A. Lacis, J. E. Hansen, P. Lee, T. Mitchell, and S. Lebedeff, 1981, "Greenhouse Effect of Trace Gases, 1970–1980," *Geophysical Research Letters* **8:**1035–1038.)

Figure 2-7. Equilibrium greenhouse warmings for estimated 1970-1980 abundance increases of several trace gases, based on climate model with sensitivity -3EC for doubled $C0_2$. (After A. Lacis, J. E. Hansen, P. Lee, I Mitchell, and S. Lebedeff, 1981, "Greenhouse Effect of Trace Gases, 1970-1980," Geophysical/ *Research* Letters **8:**1035-1038.)

About two-thirds of the chlorofluorocarbon increase for 1880-1980 occurred in the 1970s, and it seems likely that the decadel rate of increase of methane was also at a maximum in the last decade.

The eventual warming for these gases added during the 1970s is about 0.2EC if the climate sensitivity is near 2EC for doubled $C0_2$, but almost 0.4EC if the sensitivity is near 4EC. However, because of the ocean's thermal inertia only some fraction, at most about half, of the warming would be expected to have appeared by the end of the decade.

Natural fluctuations of the smoothed global mean temperature are of the order of 0.1EC for decadel periods. For example, the standard deviation of the 5 year smoothed global temperature in Figure 2-1 is 0.1EC for 10 year intervals. Therefore, although the observed warming in the 1970s (Figure 2-1) is consistent with the increased trace gas abundances, the changes cannot be confidently ascribed to the greenhouse effect.

However, if the abundance of the greenhouse gases continue to increase with at least the rate of the 1970s, their impact on global temperature may soon begin to rise above the noise level. For such a rate of increase the total warming at equilibrium due to gases added in the 1970s and 1980s would be about 0.5EC, for a climate sensitivity of 3EC. Moreover, one would expect that for a 20 year period, a large part of the equilibrium warming would appear by the end of the period. This possible warming should be compared to the standard deviation of observed temperatures, which is about 0.15EC for a 20 year period. This comparison is the basis for anticipating that significant warming is likely to occur by 1990, raising the mean global temperature well above the maximum of the late 1930s (Lacis et al., 1981).

A more quantitative statement about the near-term climate effects of increasing greenhouse gases requires a better understanding of transport and storage of heat in the ocean. This includes the transient response of the ocean to the slowly changing heating pattern at the ocean surface. Realistic treatment will require consideration of the full three-dimensional structure of ocean and atmospheric transports. It will be particularly important to determine the effect of warming and other climate changes at the ocean surface on the ocean mixing and circulation, and thus the ocean feedback on the climate change.

SUMMARY

We conclude that there is strong evidence that a doubling of atmospheric $C0_2$ will lead to a global warming of at least 1.5EC. Almost all projections of atmospheric composition indicate that an effective doubling Of $C0_2$, including contributions of trace gases, will occur sometime in the next century. Furthermore, for any climate sensitivity in the range 3±1.5EC, the global mean warming should exceed natural climate variability during the next one to two decades.

We are left in the very unsatisfactory position of having clear evidence that important climate effects are imminent but not having the knowledge or tools to specify these effects accurately. The principal areas of uncertainty include the equilibrium climate sensitivity, especially the contribution of clouds, and the nature of the transient climate response, which depends on storage and transport of heat by the ocean, including the feedbacks that may occur with changing climate at the ocean surface.

Studies of these components of the climate system are thus suggested as a high priority for research. The chief needs are observational, both global monitoring and local measurements of processes. However, to be effective, such observations must be guided by theoretical studies and modeling. It is particularly important that climate models be developed to reliably simulate regional climate, including the transient response to slowly changing atmospheric composition. This will be difficult because the models need to realistically simulate the effect of greenhouse warming on such factors as standing and transient atmospheric longwaves and ocean currents. The development of such modeling capability will take substantial time and effort, but the benefits from improved understanding of future climate effects will surely warrant the work invested.

NOTES

1. The mean radiating level can be estimated as the average altitude at which the optical path length of emitted radiation (J_L/μ) is unity for the mean cosine of emission angle μ - 0.5, with the average over frequencies L weighted according to the Planck function for the temperature at the emitting level. For Earth the global mean lapse rate is ' - 5.5EC km^{-1}, and based on the spectral computations of thermal emission, the mean altitude of emission to space is H - 6 km.
2. See note 1.

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