Observational determination of the greenhouse effect

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Satellite measurements are used to quantify the atmospheric greenhouse effect, defined here as the infrared radiation energy trapped by atmospheric gases and clouds. The greenhouse effect is found to increase significantly with sea surface temperature. The rate of increase gives compelling evidence for the positive feedback between surface temperature, water vapour and the greenhouse effect; the magnitude of the feedback is consistent with that predicted by climate models. This study demonstrates an effective method for directly monitoring, from space, future changes in the greenhouse effect.

IN its normal state, the Earth–atmosphere system absorbs solar radiation and maintains global energy balance by re-radiating this energy to space as infrared or longwave radiation. The intervening atmosphere absorbs and emits the longwave radiation, but as the atmosphere is colder than the surface, it absorbs more energy than it emits upward to space. The energy that escapes to space is significantly smaller than that emitted by the surface. The difference, the energy trapped in the atmosphere, is popularly referred to as the greenhouse effect, G. Some very interesting questions about the greenhouse effect involve feedbacks between the various elements of the climate system[1]. These feedbacks tend to enhance or diminish small external perturbations in the radiation balance. One of the most important is the water-vapour feedback[2]. Consider an initial increase in the global temperature because of the direct infrared trapping resulting from, say, an increase in carbon dioxide. As the surface and the atmosphere warm, the saturation vapour pressure increases exponentially with temperature, and more H_2O can evaporate into the atmosphere. Because water vapour is a strong greenhouse gas, the extra water vapour traps more longwave radiation and drives the temperature even higher. This feedback can amplify the original perturbation by as much as a factor of two[1,2].

The greenhouse effect and its feedback on the climate have been explored by hundreds of theoretical and model studies. Satellite radiation-budget measurements have also been used to examine the radiative feedbacks in the climate system[3]. Yet to our knowledge the radiative trapping by the atmosphere has not been examined in depth with observed data (with the exception of unpublished work by R. D. Cess). Here we address three critical issues: (1) the magnitude of G in the present atmosphere; (2) the quantitative nature of the feedback between surface temperature and G; (3) the change in G with time. We provide observational estimates for the former two and discuss how our findings bear on the third issue. A possible change in G is of urgent concern because of the rate at which human activities have been adding greenhouse gases to the atmosphere.

We obtain G by subtracting longwave radiation escaping to space from estimates of the radiation emitted by the ocean surface. The radiation escaping to space has been measured since 1985 by the spaceborne Earth Radiation Budget Experiment (ERBE). Under cloudless skies, the atmosphere traps 146 W m^{-2} whereas clouds increase G by 33 W m^{-2}. G increases significantly with surface temperature (T_s), even when it is normalized by the surface emission. The rate of increase of G, 3.3 W m^{-2} K^{-1}, is consistent with the magnitude of the H_2O greenhouse feedback inferred from climate models and line-by-line radiative-transfer calculations. Furthermore, the observed variation of total atmospheric water content (W) with T_s indicates that the normalized greenhouse effect (g) is strongly correlated with W. The observations are compelling evidence for the H_2O feedback. The data also reveal a rapid rise in G at the higher temperatures ($T_s > 298$ K), the cause(s) of which are not apparent.

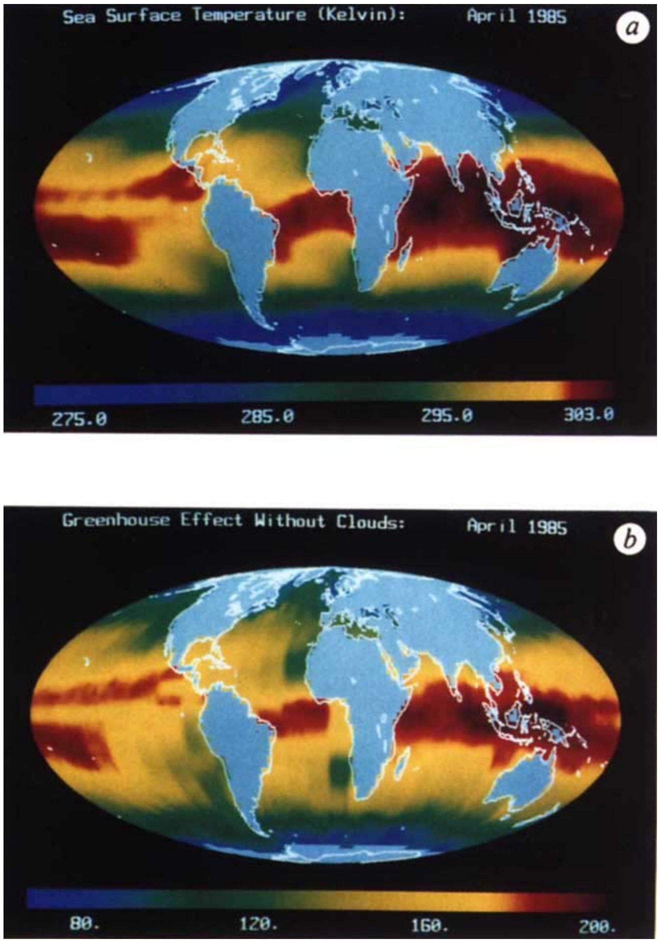

FIG. 1 *a*, Monthly-mean map of observed sea surface temperature (K) for April 1985. *b*, Same as in *a* but for the greenhouse effect (G) of atmosphere under clear skies. Values are in W m^{-2}. The major inference from these figures is that the greenhouse effect of the atmosphere is strongly correlated with surface temperature. This correlation can explain the Equator to pole decrease in G as well as the east to west variations. For example, G is maximum over the relatively warm western equatorial Pacific Ocean (Indonesian region) and is very low over the cold oceanic upwelling regions in the central and eastern equatorial Pacific and off the west coasts of Africa and South America.

Determination of the greenhouse effect

If E is the longwave flux emitted by the surface at a certain location, and F is the flux leaving the top of the atmosphere (TOA) directly above that location, then the greenhouse effect G for that location can be defined as $G = E - F$.

The TOA fluxes F have been measured with good accuracy by scanning radiometers on board the spaceborne ERBE[4]. ERBE consists of three satellites carrying identical scanners that detect both longwave (infrared) and shortwave (solar) radiation. The ERBE scanners, when viewing directly down, can resolve an area of ~35×35-km. The scanners have in-flight calibration which assures an accuracy of 1% for the satellite-altitude readings. Those readings are converted to TOA fluxes on a global grid of 2.5°×2.5° (latitude×longitude). So far, we have processed data for the months of April, July and October 1985 and January 1986. A unique feature of ERBE is that it separates the fluxes for clear skies (cloudless conditions) from fluxes for cloudy skies (mixture of clear and overcast areas)[5]. Hence we can obtain the greenhouse effect of the atmosphere and that of clouds separately.

We estimate the emission from the surface using the Stefan-Boltzmann black-body law ($E = \sigma T^4$) and the observed surface temperature. Because, in the infrared, the ocean surface approximates a blackbody to within 1% (ref. 6), and because the diurnal surface temperature variation is small in the ocean's mixed layer (less than a few degrees K), the uncertainties in the calculated values of E are minimum over oceans. We restrict this study to the open oceans; that is, we ignore continents and ocean regions covered with sea ice.

We obtained surface temperatures (SSTs) from the National Meteorological Center daily data sets archived at the National Center for Atmospheric Research (NCAR). The SST values are derived by blending *in situ* (ship and buoy) data with satellite temperature retrievals[7].

The SSTs used during the period of this study have an estimated error of about ±1.0 K (ref. 7). The error in the surface emitted flux is simply $\Delta E = 4\sigma T^3 \Delta T$, or about ±5 W m^{-2}. ERBE TOA longwave fluxes have an estimated error of about ±5 W m^{-2} for monthly averaged values[4,5]. Because the errors in E and F are uncorrelated, the error in the monthly averaged greenhouse effect for a single location is ±7 W m^{-2}. When these values are averaged around a latitude belt, the random error will diminish significantly. Clear-sky fluxes, however, can have systematic errors that are due to cloud contamination. The magnitude of this error was examined[5] by sorting out homogeneous cloudless regions, that is cloudless regions that are surrounded by other cloudless regions and for which the spatial standard deviation of the longwave flux is <1% of the clear-sky flux. This stringent homogeneity criterion gave fluxes that were smaller than the operational values used here by only 4 W m^{-2}, indicating minimal cloud contamination[5]. In addition, the clear-sky fluxes when compared with state-of-the-art radiation models agreed within 1% (ref. 5). Overall, we expect the error (systematic plus random) in the monthly and regional mean values of G to be 5 to 10 W m^{-2}.

Global greenhouse effect of atmosphere and clouds

Estimates of G over the open oceans for a cloudless atmosphere (clear sky) reveal significant regional variations (Fig. 1*b*). Globally, for April 1985, the surface emission is 421 W m^{-2} whereas the TOA emission is 243 W m^{-2}, resulting in a total trapping of 178 W m^{-2}. Of this, the atmospheric gases (H_2O, CO_2 and O_3 among others) trap 146 W m^{-2}, whereas clouds trap the remaining 33 W m^{-2}. The greenhouse effect for each of the months studied is given in Table 1.

As the months are taken from the four seasons, we will assume the four month mean to be representative of the annual average. Then the annual average value of G (179 W m^{-2}) is comparable to the global mean solar absorption, ~237 W m^{-2}. The global mean latent heat released in the atmosphere is ~100 W m^{-2}. The greenhouse effect of a doubling of CO_2 is 4 W m^{-2} and that of human activities during the past century[8] is ~2 W m^{-2}. The greenhouse effect of the atmosphere, next to solar absorption, is the largest factor in maintaining the present climate.

The TOA flux is given by the radiation transfer equation

$$F = B(T_s) - \int_0^1 A(x)[\mathrm{d}B/\mathrm{d}x]\,\mathrm{d}x \tag{1}$$

This equation is written in a normalized pressure coordinate, $x = p/p_s$, where p_s is the surface pressure. The blackbody emission B is simply $\sigma T^4 = \int B_\lambda(T)\,\mathrm{d}\lambda$, where B_λ is the monochromatic Planck function evaluated at atmospheric temperature T. The effective absorptivity A is the integral of the monochromatic absorptivity A_λ weighted with $\mathrm{d}B_\lambda$, and normalized by $\mathrm{d}B$ (ref. 2). Thus $A(x)$ is the absorptivity between the TOA ($x = 0.0$) and the pressure level x. T_s is the surface temperature. Because $G = E - F$ and $E = B(T_s)$, we obtain

$$G = 4\sigma \int_0^1 A(x) T^3 [\mathrm{d}T/\mathrm{d}x]\,\mathrm{d}x \tag{2}$$

where we have let $(\mathrm{d}B/\mathrm{d}x) = 4\sigma T^3(\mathrm{d}T/\mathrm{d}x)$. Most of the contribution to A comes from the troposphere (the bottom 10 km) which contains the bulk of the absorber mass. (Although CO_2 is uniformly mixed in the atmosphere, the major absorbers—water vapour and clouds—are almost entirely restricted to the troposphere.) In the troposphere temperature generally decreases with height. Hence $(\mathrm{d}T/\mathrm{d}x) > 0.0$ and G is positive. This reveals that for G to be positive, the temperature must decrease with altitude in the region of the absorbing gas.

Evidence of water vapour feedback

The clear-sky G (Fig. 1*b*) generally decreases from Equator to pole with some exceptions. The trapping that is due to clouds (not shown here) does not vary systematically with latitude, but peaks in the convectively disturbed regions of low latitudes and in the vicinity of the mid-latitude jet-stream regions. We find that the clear-sky G has a strong positive correlation with sea surface temperature (compare Fig. 1*a* with *b*). Here we show that this correlation gives direct evidence of the H_2O feedback.

We do this in three steps. First, we show the correlation of G with T_s implies that the longwave absorptivity of the atmosphere increases with surface temperature. Second, the variation of water vapour in the atmosphere is determined largely by the increase in the saturation vapour pressure with temperature, as shown by other observational[2,9] and modelling[10] studies. Third, if the observed variation of water vapour causes the sort of G-T_s coupling revealed by the data, the atmosphere absorptivity must scale logarithmically with the water content. This is not only indicated by the observations but is also consistent with theories of radiative transfer.

Step 1. Because the surface emission increases as the fourth power of T_s, the trapping should also increase with T_s. To remove the T_s^4 dependence, we define a normalized trapping $g = G/\sigma T_s^4$. Both the clear-sky g (Fig. 2) and the cloudy g (not shown) increase nonlinearly at warm temperatures ($T_s > 298$ K), but otherwise g increases smoothly with T_s. The grid-point

TABLE 1 Monthly- and global-mean greenhouse effect (W m^{-2})

Month	Clear sky	Clouds	Cloudy sky*
April 1985	146.3	32.1	178.4
July 1985	147.7	31.3	179.0
October 1985	146.9	32.9	179.8
January 1986	143.7	34.1	177.8
Four-month mean	146	33	179

* Cloudy sky is the sum of clear sky and clouds.

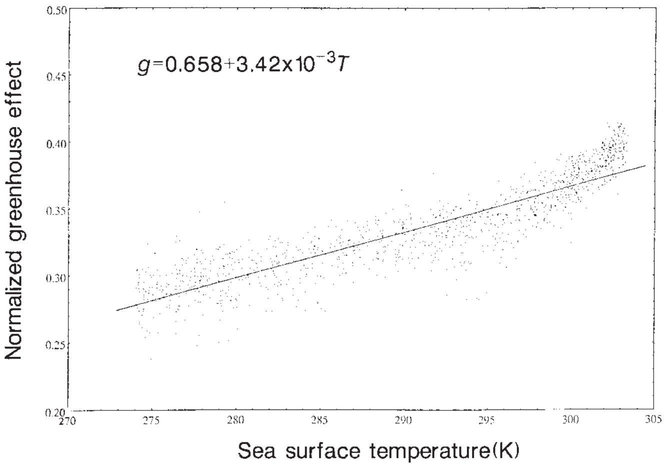

FIG. 2 Clear-sky greenhouse effect normalized by infrared emission (g) as a function of surface temperature, April 1985. Values are monthly averages for each 2.5° × 2.5° region. The solid line indicates a least-squares fit through all the points. Fits done individually through Northern and Southern Hemisphere points do not deviate by more than 3%.

values of g for the individual months and the two hemispheres reveal similar dependencies on T_s. Because of the robustness of this feature, we conclude that g and the surface temperatures are strongly coupled.

The increase in g is primarily because of the increase in the absorptivity with T_s. In the troposphere the lapse rate $\Gamma = -dT/dz$ is approximately constant with pressure. We can then express the atmospheric temperature as $(T/T_s) = x^{\alpha}$, where $\alpha = \Gamma R/g_a$, where g_a is the gravitational acceleration. Inserting this in equation (2)

$$g = G/\sigma T_s^4 = 4\alpha \int_0^1 A(x) x^{\beta}\, dx \qquad (3)$$

where $\beta = (4\alpha - 1)$. We first see that g does not explicitly depend on T_s. The g-T_s coupling must arise because either the lapse rate (and hence α) or A depend on T_s. The observed change in lapse rate[11] over the range of T_s considered here can affect g by at most 10% (based on flux sensitivities to lapse rates given in ref. 12). The only other variable is A. For a fixed amount of H_2O, A varies by only 5% (ref. 2) for the range of T_s considered here. The temperature dependence of A also cannot explain the rate of increase of g with T_s. Therefore, the amount of water vapour must increase with surface temperature.

Step 2. The increase in water vapour with T_s, which in turn causes an increase in A, is sufficient to explain the g-T_s coupling. The saturation vapour pressure e_s of H_2O is given by the Clausius-Clapeyron equation $C\exp\{-(LR^{-1})/T\}$. Here, C is a constant, L is the latent heat of vaporization and R is the gas constant for water vapour; LR^{-1}, at the global mean temperature, is ~5,400 K. Using the observed global mean lapse rate and relative humidity profile[2] we numerically calculate the total water content W in a vertical column in the troposphere for a range of values of T_s. W increases exponentially with a rate of increase $d(\ln W)/dT_s = 6.7\times10^{-2}\ \mathrm{K}^{-1}$. Compare this to the actual rate of increase ($5.5\times10^{-2}\ \mathrm{K}^{-1}$), obtained by satellite microwave observations[13] of W (Fig. 3a). They agree to within 20%, hence W is largely determined by the surface temperature, in agreement with the findings of Prabhakara *et al.*[9] and Stephens[14]. This coupling between W and T_s arises from the temperature dependence of the saturation vapour pressure.

Step 3. Because we are ignoring variations in the lapse rate or in spectroscopic parameters with temperature, we can write dg/dT_s as a product of a partial and a total derivative:

$$\frac{dg}{dT_s} = \frac{\partial g}{\partial \ln W} \cdot \frac{d\ln W}{dT_s} \qquad (4)$$

The first term is the observed coupling of the normalized greenhouse effect with T_s. The last term represents the fundamental source of the H_2O feedback—the water content increases with T_s. The second term represents the 'return loop' in the feedback—the absorptivity increases with water content. We infer from equation (4) that if g varies linearly with T_s (which it does for $T_s < 298$ K in Fig. 2), and if ln W also varies linearly with T_s (as suggested from theory and from the satellite microwave data in Fig. 3a), then g should increase logarithmically with W. In fact, theoretical calculations[2] (see ref. 15) have shown that, for the range of W considered in Fig. 3, the absorptivity scales logarithmically with the absorber amount. A plot of g against ln W (Fig. 3b) shows this scaling. Our analysis is consistent, because dg/dT_s (from Fig. 2) is $3.4\times10^{-3}\ \mathrm{K}^{-1}$, on a global average, whereas the product of $\partial g/\partial(\ln W)$ (from Fig. 3b) and $d(\ln W)/dT_s$ (from Fig. 3a) is $3.2\times10^{-3}\ \mathrm{K}^{-1}$. This comparison may be used in an illustrative sense only, because the microwave soundings of W are only available for the period 1979–83, and the ERBE measurements are not available before 1985. However, a comparison of zonal averages (as shown in the plot of g against ln W in Fig. 3b) minimizes the variation across these two time periods.

The analysis strongly indicates the presence of the H_2O feedback. Three further points can be made in favour of this interpretation. First, the clear-sky greenhouse effect and its temperature

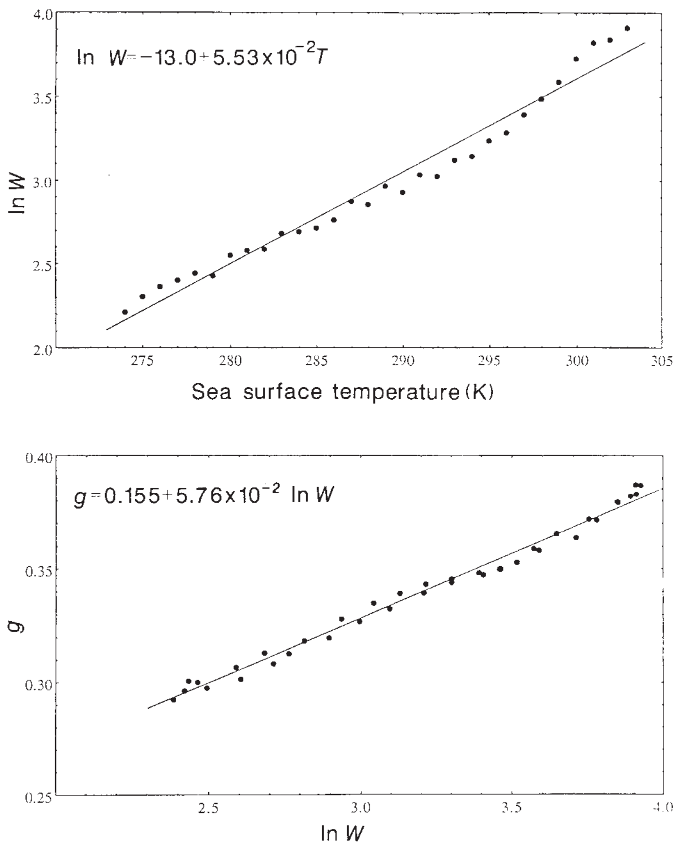

FIG. 3 a, Natural logarithm of the atmospheric water content (W, kg m^{-2}) as a function of surface temperature, April 1982, monthly averages for 3° × 5° regions. The water content was derived from microwave satellite soundings[13] available for 1979–1983. The error in W is ~10% (ref. 13). The strong positive correlation indicates that the behaviour of W follows simple thermodynamic laws; the slight upward and then downward deviations are consistent with the latitudinal variations in relative humidity[2] and lapse rate[11] which are governed by the dynamics of the atmosphere. b, Normalized clear-sky greenhouse effect (g) for April 1985 plotted against ln W for April 1982. ERBE data is not available before 1985 whereas W is not available after 1983. To minimize the effect of year to year variations in W, we have used zonally averaged values.

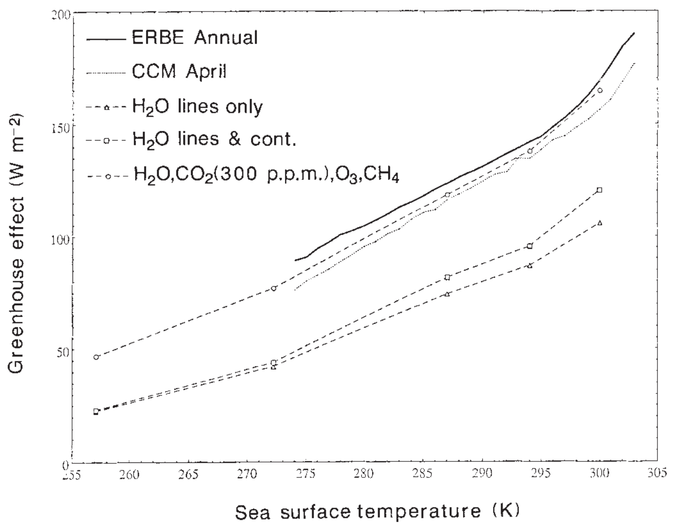

FIG. 4 Comparison of G and surface temperature, retrieved from three sources: bold line, ERBE annual values, obtained by averaging April, July and October 1985 and January 1986; thin line, three-dimensional climate-model simulations for a perpetual April (National Center for Atmospheric Research, Community Climate Model); dashed lines, line-by-line radiation-model calculations by Dr A. Arking (cited in ref. 16) for individual H_2O spectral lines, as well as the continuum absorption. The five points use climatological winter and summer atmospheric profiles for subarctic and mid-latitude conditions, and a tropical profile. A comparison of the calculated G with and without the continuum absorption (bottom two dashed lines) indicates a nonlinear contribution from the continuum. This may in part account for the nonlinear rise at the highest temperatures in the ERBE-derived G. When CO_2, O_3 and CH_4 are added (the top dashed line) the line-by-line model results come close to ERBE values.

dependence is consistent with state-of-the-art radiative-transfer calculations[16] (Fig. 4). These calculations use observed humidity and temperature profiles for each climate type (subpolar winter, midlatitude summer, for example). Hence, their agreement with the observed G indicates that the 'return loop' in the H_2O feedback, $\partial g/\partial(\ln W)$, behaves as expected from theory. Second, the sensitivity dG/dT_s of the NCAR Community Climate Model (CCM)[17] agrees closely with the observed sensitivity (Fig. 4). The CCM is a three-dimensional general circulation model. The model ignores the radiative effects of trace gases such as CH_4 and N_2O (which contribute ~5-7 W m^{-2} to the clear-sky G) and is also drier than the observed climate[18]. Hence it systematically underestimates G by 10-15 W m^{-2}. It has, however, a latitudinal sensitivity dG/dT_s of 3.1 W m^{-2} K^{-1}, which compares well with the value of 3.3 W m^{-2} K^{-1} obtained from ERBE, averaged annually and over the whole range of sea surface temperatures.

Finally, the general circulation model (GCM) designed by S. Manabe shows a climate sensitivity consistent with the analysis of ERBE flux observations. Wetherald and Manabe calculated[19] that a 2% increase in the solar constant increased the GCMs outgoing longwave flux by 6.3 W m^{-2}, and the surface temperature increased by 3.0 K. Thus the longwave trapping of the GCM has a sensitivity to the global mean temperature of $dG/dT_s = 4\sigma T_s^3 - dF/dT_s$, or 3.7 W m^{-2} K^{-1}. A 2% decrease in the solar constant results in a higher sensitivity, 4.0 W m^{-2} K^{-1}. The cloud cover is held constant in this model, hence the cloud-cover feedback is zero. The observed sensitivity is within 15% of the GCM values. This implies that the climatological sensitivity of this GCM is in good agreement with the latitudinal sensitivity of the present atmosphere. Such behaviour is necessary if the change in the equilibrium climate is determined largely by the H_2O greenhouse feedback.

The analysis up to this point is very encouraging—the independent climate data sets show an overall consistency that can be explained in terms of thermodynamics and radiation physics. This is perhaps one of the best indicators of the accuracy of satellite data. We find compelling evidence that the H_2O feedback is detectable from ERBE observations and that it can account for most (at least 80%) of the observed variation of G with surface temperature. Other atmospheric processes, such as the variation of relative humidity or lapse rate with latitude, are needed to explain the remaining variation.

Implications to the global warming problem

The coupling between surface temperature and the greenhouse effect will lead to a strong positive feedback on any perturbations to the present climate. One significant implication is that we might detect the greenhouse effect of human activities. The increase in trace gases, if it continues unabated, will directly increase G by about 0.5 W m^{-2} per decade. The cumulative increase in G for the next 50 years will then be ~2.5 W m^{-2}. In the same period, the globe is expected to warm by 2-4 K (ref. 8). Because of the coupling between T_s and g, the global G will actually increase by a further 6.5-13 W m^{-2}, leading to a total increase of 9-15.5 W m^{-2}. Such an increase should be clearly detectable if we continue ERBE-type observations into the next century. If the tropical regions exhibit the nonlinear increase revealed by the ERBE data (Fig. 2), the effect should be most prominent in the tropics.

Both the cloudy and clear-sky greenhouse effect increases rapidly at temperatures characteristic of the tropics. It is at present not known whether this nonlinear rise is simply a climatological feature of low latitudes, unrelated to temperature variations. For example, the relatively dry regions found in the subtropics ($290 < T_s < 298$ K) may slightly suppress the rate of increase of g with T_s and hence change the slope.

The nonlinearity might, on the other hand, be due to the nonlinear dependence of the H_2O continuum opacity[16] or a temperature threshold for deep convection[20]. Because deep convection is a source for the ice content of cirrus clouds, the greenhouse effect of cirrus clouds can increase with T_s when T_s exceeds a certain threshold value. If so, the magnitude of the positive feedback is so large that we should look for a compensating negative feedback elsewhere in the system to explain the relative stability of the tropical climate during the last great ice age. □

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