SIMPLE CLIMATE MODELING

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ABSTRACT

Climate models are tools for scientific and policy analysis. Two simple energy balance models (EBM) of the climate of the earth were built for sensitivity studies. Using these models, we demonstrate that: changes in the albedo of the atmospheric-surface system contribute to radiative damping, solar variability is an important contributor to global mean temperature changes; a projection of future global temperatures reveal that by 2100, almost 50% of the greenhouse gas forcing will be due to non-CO_2 greenhouse gases; the projection of global mean temperature changes shows an estimated increase by 2100 between 1 and 2 C; if the greenhouse gases are reduced by 5% (Kyoto Protocol agreement that levels at 2010 be reduced to 1990 levels), we estimate the mean temperature to decrease by 0.5 C by 2100. Evaluation of the one-dimensional EBM show that the simulated values are not significantly different from observed values at the 99% confidence value.

INTRODUCTION

The enhancement of anthropogenic emissions of greenhouse gases (GHG) into the atmosphere is leading to human induced climate changes that are likely to have important impacts on natural and human systems. As a response to this emerging threat, the Kyoto Protocol[1] was signed in 1997, committing developed countries to a 5% reduction of the emissions of a ‘basket’ of 6 GHGs (to 1990 levels) by the commitment period 2008-2012. To predict the behavior of the climate on variables such as GHGs, climate models are developed. Despite being a young discipline, not yet 30 years old, climate modelers have a great responsibility thrust upon them by the ratification of the United Nations Framework Convention on Climate Change. [2] The objective of this convention is to achieve stabilization of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system. [3]

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Our ability to project and detect future climate change is limited by uncertainties regarding:

- ° estimation of future emissions and biogeochemical cycling (including sources and sinks) of greenhouse gases, aerosols, aerosol precursors, projections of future concentrations of the gases and aerosols, and radiative properties of the gases and aerosols;
- ° representation of climate processes in models, especially feedback associated with clouds, oceans, sea-ice and vegetation;
- ° systematic collection of long-term instrumental observations of climate system variables (such as solar output, atmospheric energy balance components, hydrological cycles, ocean characteristics and ecosystem changes) for the purpose of model testing, assessment of temporal and regional variability, and for the detection and attribution studies.

To quantify the response of the climate to changes in forcing, it is essential to account for all complex interactions and feedbacks among the various climate system components. It is not possible to do this reliably using empirical or statistical models, due to the complexity of the system and because the possible outcomes may go well beyond any conditions ever experienced previously. Instead, the response must be found using numerical models of the climate system based upon sound well-established physical principles.

This paper focuses on the study of two energy balance

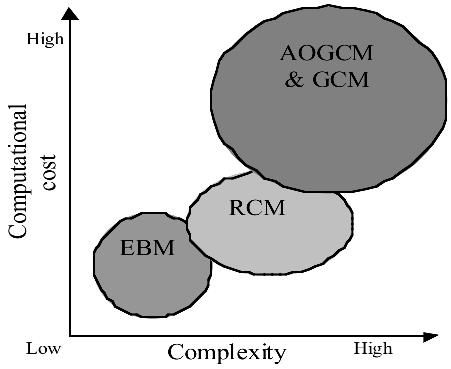

Figure 1
Schematic comparison [4] between different climate models in terms of computational costs and complexity.

models (EBMs), their creation, results from the models and interpretations of the results. The EBMs allow us to predict surface temperature variations with latitude using simplified relationships based on sound physical principles. These models have been instrumental in increasing our understanding of the climate system and in the development of new parameterization methods for evaluating sensitivity for more complex and realistic models. In climate modeling, parameterization is defined as the method of incorporating a process by representation as a simplified function of some other fully resolved variables without explicitly considering the details of the process. [2]

Figure 1 compares various climate models in terms of computational cost and complexity. Computation cost is determined by the computing facilities needed to run the model: personal computer (low) or supercomputer (high). Complexity is the level of detail with which the individual model components are treated. We chose EBMs due to their low complexity and computational costs.

There are more complex and computationally demanding models: circulation models (GCM) and coupled atmospheric-radiative convective models (RCMs), general ocean general circulation models (AOGCM). These complex models can simulate past and present geographical variation of temperature as well as other climatic variables such as rainfall, evaporation, soil moisture, cloudiness and winds. [4] These models provide credible continental scale changes of some of the variables. The use of these models is limited by their high computational costs.

THE MODELS

JavaScript was used to create the models we used. They are available on-line. (http://www.uea.ac.uk/~x9723668/scm/models/model0.html) The first model is a zero-dimensional energy balance model that considers the earth as a single point in space, having a global mean surface temperature of T_S. Viewing the earth from the outside, one observes an amount of radiation input which is balanced in the long term by an amount of radiation output by the planet (i.e., the system is in equilibrium). Since over 70% of the energy which drives the climate system is first absorbed at the surface, the surface albedo will be dominant in controlling the energy input into the climate system. [2]

If α is the average planetary albedo (the reflected fraction of incident radiation), the power absorbed by the earth will be:

$$E_{in} = (1 - \alpha)\, S\, \pi R^2, \qquad (1)$$

where S is the solar constant (here considered to be 1370 Wm^{-2}) and R is the radius of the earth. If the earth is assumed to be a blackbody, the power emitted by the earth is given by the Stefan Boltzmann law:

$$E_{out} = 4\pi R^2\, \sigma\, T^4\,, \qquad (2)$$

where T is the temperature of the earth and σ is the Stefan-Boltzmann constant ($5.67 \times 10^{-8}\ Wm^{-2}K^{-4}$). For the earth to be in equilibrium, the incoming power (Equation 1) must equal the outgoing terrestrial power (Equation 2). Setting Equation 1 equal to Equation 2 gives:

$$(1 - \alpha)\, S\, \pi R^2 = 4\pi R^2\, \sigma\, T^4\,. \qquad (3)$$

Solving Equation 3 for the radiation balance temperature T, and adding an increment ΔT to model the GHC contribution to the surface temperature, the surface temperature of the earth T_S can be written as:

$$T_s = \sqrt[4]{\frac{(1-\alpha)S}{4\sigma}} + \Delta T. \qquad (4)$$

In our modeling, we assume a value of 33 K for ΔT. Using the model outlined in Equation 4, it is possible to observe the effect of the albedo on the surface temperature of the earth.

A second more complex one dimensional EBM considers each latitude zone of the earth independently. Dividing the earth into latitudes provides a more realistic approach than

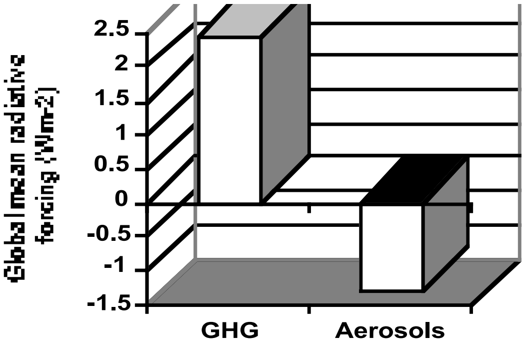

Figure 2
Estimates of the globally and annually averaged anthropogenic radiative forcing due to changed in concentrations of greenhouse gases and aerosols from pre-industrial times to the present . [8]

the first model which only supplies the global mean surface temperature. This model can be accessed at http://www.uea.ac.uk/~x9723668/scm/models/model1.html. The model is controlled by three main relationships:
the radiation entering the earth at latitude ϕ is given by:

$$R_{in}(\phi) = \frac{S}{4}\, SunWt(\phi)\ , \qquad (5)$$

where $SunWt(\phi)$ is the solar radiation distribution at each latitude;
the surface temperature at latitudes ϕ is given by

$$T_s(\phi) = \frac{R_{in}(\phi)[1-\alpha(\phi)] + (K)\,(GMT) - A}{B+K}, \qquad (6)$$

where $\alpha(\phi)$ is the initial albedo at latitude ϕ, GMT is the global mean temperature (assumed to be 14.87 C) and K,A and B are empirical constants.

The outgoing radiation at latitude ϕ is given by:

$$R_{out}(\phi) = A\ + (B)T_s(\phi) + (\alpha)R_{in}(\phi). \qquad (7)$$

The first part of Equation 7 is the long-wave radiation from the sky to space and the second part is the reflected incoming solar radiation. The albedo of the earth below 70˚ latitude is assumed to be 0.3, and the ice above 70˚ latitude will increase the albedo to 0.6.

Definition of terms used in climate modeling
An important aspect of models is 'radiative forcing'. Radiative forcing is the perturbation of the energy balance of the surface-troposphere system after allowing the stratosphere to readjust to a state of global mean radiative equilibrium. [4,5] An example of radiative forcing is the perturbation of the energy balance following a change in the concentration of carbon dioxide or a change in the output of the sun. A positive radiative forcing tends to warm the surface, while a negative forcing tends to cool it. Some external forcings are positive forcing by greenhouse gasses (CO_2, CH_4, N_2O, CFC-11, CFC-12 HCFC22 and other halocarbons) at a rate of 2.45 Wm^{-2} and an estimated (within a factor of 2) of negative forcing by aerosols of -1.3 Wm^{-2}. Figure 2 shows the global mean radiative forcing due to two mechanisms, GHG and aerosol forcing, since pre-industrial times. These are the only two mechanisms that have been considered with sufficient confidence to matter in the modeling processes.

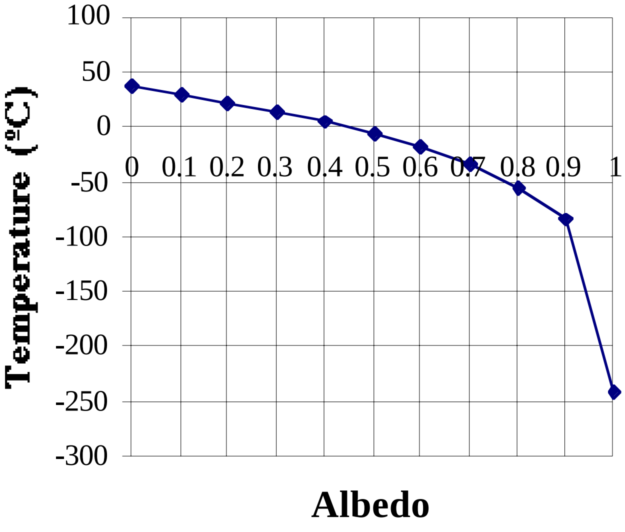

Figure 3
Global mean surface temperature change vs albedo using a zero-dimensional energy balance model.

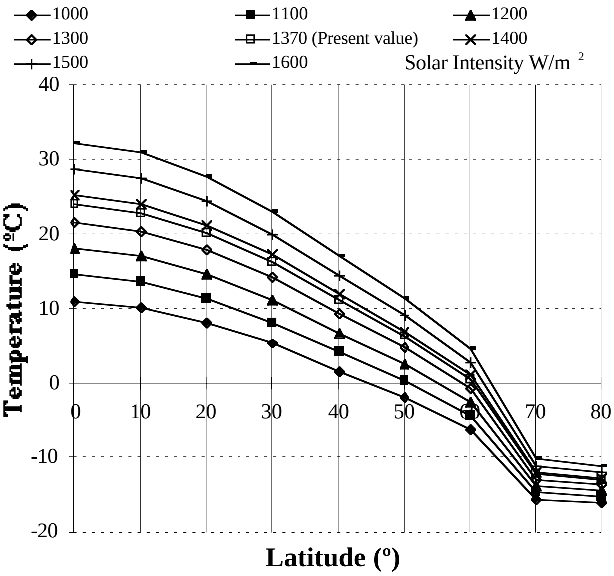

Figure 4
Surface temperature at different latitudes for different solar energy using a one dimensional energy balance model.

'Climate sensitivity' is the steady-state increase in the global annual mean surface air temperature associated with a given global mean radiative forcing. [4] It is common practice to use CO_2 doubling as a benchmark for comparing climate model sensitivities.

The 'feedback term' parameterizes the effects the processes involving water vapor, clouds, ice and snow, and ocean-atmosphere dynamics. [6] In equilibrium,

$$\Delta Q = \lambda\, \Delta T, \qquad (8)$$

where ΔQ represents the external forcing, ΔT is the change in global mean temperature and λ is the atmospheric feedback parameter. An example is the case where the solar absorption does not change. Then the derivative of Equation 1 can be used to find DQ:

$$\Delta Q = \frac{\partial Q}{\partial T}\,\Delta T = 4\sigma T^3\,\Delta T\ . \qquad (9)$$

Using Equation 8 and picking an effective emission temperature of 255K gives a value of the feedback term as:

$$\lambda = 4\sigma T^3 = 3.8\ \mathrm{Wm}^{-2}\ . \qquad (10)$$

This value coincides with the one that can be extrapolated by knowing that the minimum global temperature has increased between 0.3 C and 0.6 C since the 19th century [4] and using the radiative forcing shown in Figure 2. Solving

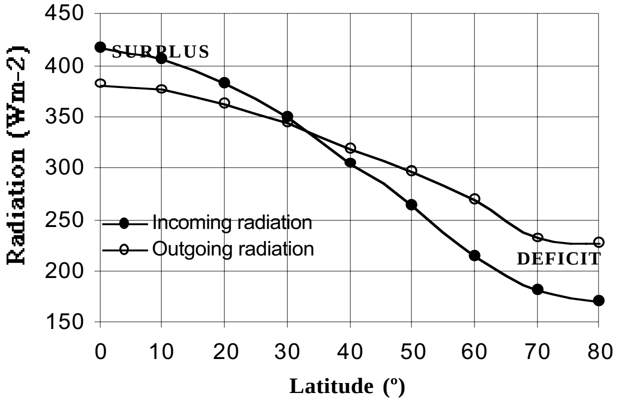

Figure 5
Radiation as a function of latitude using a one dimensional energy balance model.

Equation 8 for λ gives:

$$\lambda = \frac{2.45 - 1.30}{0.3}\ \mathrm{Wm}^{-2} = 3.8\ \mathrm{Wm}^{-2}. \tag{11}$$

For our modeling, we pick feedback terms of values 1, 2, 3.3 and 3.8 Wm^{-2} and radiative forcing values shown in Table 1. To compare the various climatological perturbations, we found an average value and then found the temperature change the perturbations had forced.

RESULTS

Results from the zero-dimensional EBM are shown in Figure 3. The line shows the temperature vs albedo. As expected, an increase in albedo leads to a decrease in the surface temperature. The earth's actual mean albedo has a value of 0.3, which produces a global mean surface temperature of about 14 C.

The one-dimensional EBM calculated values of surface temperature, incoming and outgoing radiation in each latitude zone. Figure 4 shows the surface temperature at different latitudes with increasing solar energy. A solar energy increase leads to a temperature increase, while a solar energy decrease provokes a temperature reduction. High temperatures can be seen near the equator, while low temperatures are found near the poles. The sudden decrease in temperature at a latitude of 70˚ is due to the change in albedo at this latitude. Figure 5 shows the incoming and outgoing radiation. There is a positive budget (surplus) at latitudes less than 35˚ and a deficit at larger latitudes. Figure 6 shows the net radiation balance of the earth's surface (incoming solar radiation minus long-wave radiation from sky to space) and the earth-atmosphere system (incoming solar radiation minus outgoing long-wave radiation) as a function of latitude.

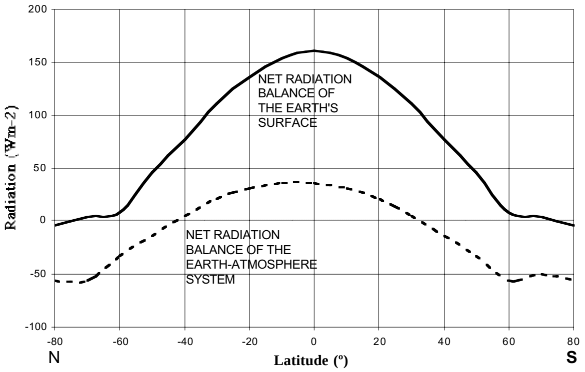

Figure 6
Net radiance balance for the surface of the earth and the whole earth-atmosphere system.

	1980	1990	1994	2100
CO_2	1.10	1.50	1.56	2.83
CO_2+GHG	*1.83*	*2.45*	*2.85*	*5.55*
GHG+aerosols	0.53	1.15	1.55	4.94
Kyoto	0.53	1.15	1.55	3.57

Table 1
Comparison of estimated radiative forcing in Wm^{-2} due to different scenarios which consider: CO_2, all GHGs, all GHGs and aerosols and the Kyoto Protocol agreement with reduction in GHGs emissions

Results of how different forcing and feedback factors effect the temperature is shown in Table 2. Figure 7 shows a comparison of the various perturbations using a mean value of λ = 2.65 Wm^{-2}, which leads to a climate sensitivity ΔT = 1.5 C (if you only consider doubling CO_2) and ΔT = 2.2 C if you consider all perturbations except clouds (due to its high uncertainty). These results have caveats because: 1) we assume that the climate system is in equilibrium; 2) we assume that the heat flux into the ocean is zero. Both of these assumptions are not true, but they simplify the modeling exercise which otherwise would not a simple task. Taking this into account, we use the best estimate [16] of ΔT = 2.5 C with λ = 1.484 Wm^{-2} and ΔQ = 3.71 Wm^{-2}. Using these values, we estimate global mean temperature change for the next century (see Figure 8) under various for various forcing combinations: CO_2; CO_2 + GHG; and GHG+aerosols and Kyoto. Notice that the CO_2 and GHG+aerosol forcing curves are in agreement

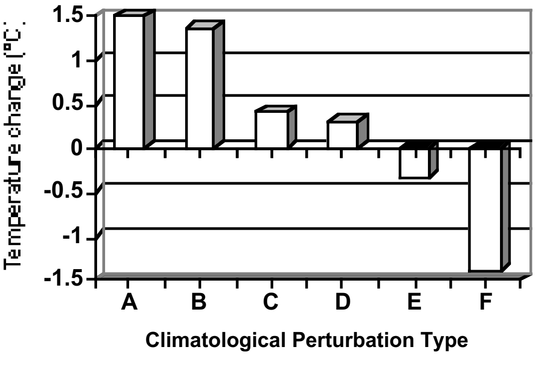

Figure 7
The effect of various climatological perturbations upon surface temperatures using a value of l = 2.65 Wm^{-2}. The perturbations legend is shown in Table 2

		λ ($Wm^{-2}K^{-1}$)			
		3.8	3.3	2.0	1.0
	ΔQ(Wm^{-2})	ΔT (C)			
A-Doubling of CO_2	3.5 : 4.0	0.9 : 1.0	1.1 : 1.2	1.8 : 2.0	3.5:4.0
B-Solar Luminosity (+1%)	3.42	0.9	1.0	1.7	3.4
C- Doubling N_2O	0.97	0.3	0.3	0.5	1.0
D-Tripling CH_4	0.70	0.2	0.2	0.4	0.7
E-Aerosols in Clouds	0.0 : -1.5	0 : -0.4	0 : -0.5	0 : -0.8	0 : -1.5
F-Volcanic Eruptions	-2 : -4	-0.5 : -1	-0.6 : -1.2	-1.0: -2.0	-2 : -4

Table 2
Temperature changes forced by climatological perturbations with different feedback factors.

with the observed global mean temperature change since 1850.

DISCUSSION OF RESULTS

EBM calculations

Figure 4 is in agreement with the fact that the average temperature in not uniform throughout the latitude zones. From Figure 4, one can deduce the changes that might have happened millions of years ago on the earth if variations in the tilt, eccentricity and precession of the earth's orbit lead to changes in the amount of solar energy striking the earth. [11] Incorporation of solar variability improves the agreement between the model and global mean temperature observations. [12]

In the radiation budget calculation shown in Figures 6 and 7, an imbalance between the equator regions and the pole regions is evident. As the tropics do not get progressively hotter or the higher latitudes colder, a redistribution of world energy must occur, probably taking the form of a continuous movement of energy from the tropics to the poles. [13] The poleward heat transport must take place within the atmosphere and the oceans.

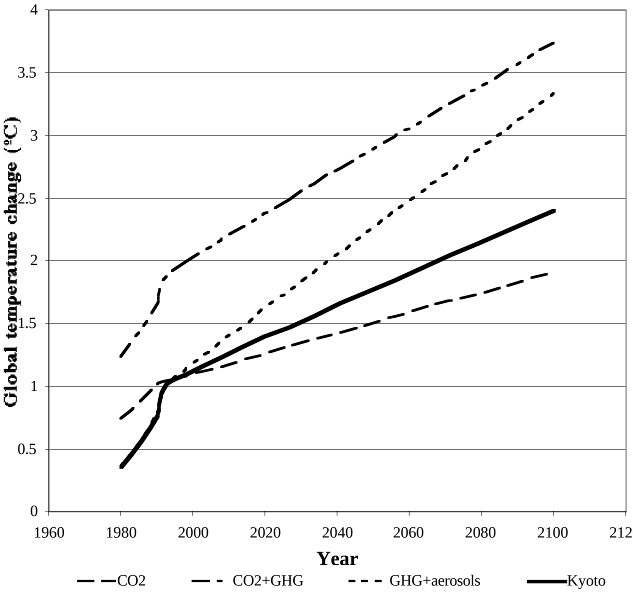

Figure 8
Global mean temperature change for different scenarios for the best estimate of climate sensitivity [11] (2.5 C) from 1980 to 2100.

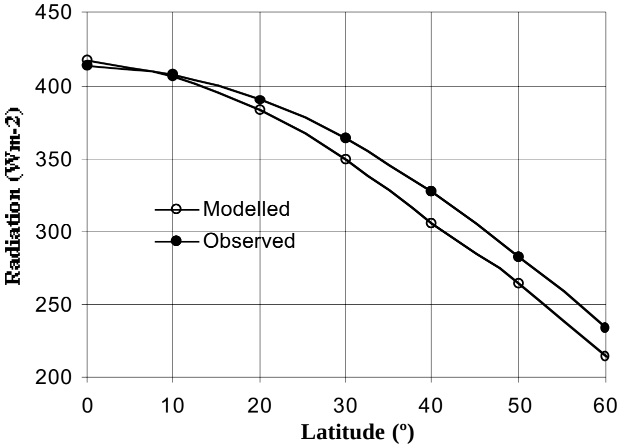

Figure 9
Observed mean short-wave radiation vs latitude [16] compared with simulated values from one dimensional energy balance model.

The reliability of the one-dimensional EBM can be seen in Figure 9, by comparing simulated and actual values. The statistical evaluation of the data shows that the simulated values are not significantly different from the observed at the 99% confidence level. The mismatch above 20° latitude is believed to be due to the energy transported from the equator polewards by the general atmospheric circulation and the oceans which is not represented by the model. Figure 10 shows the comparison of the simulated and observed long-wave radiation from the sky to space.

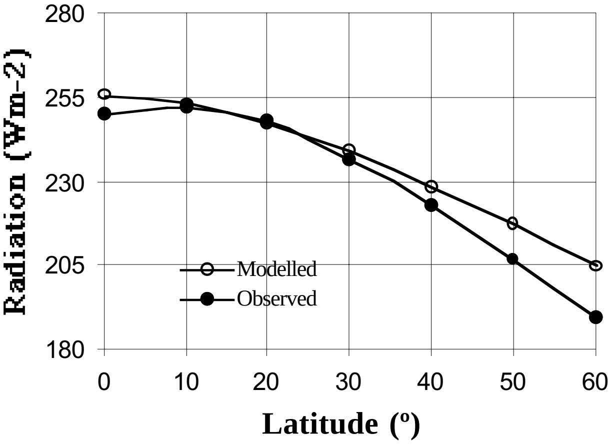

Figure 10
Observed annual mean long-wave radiation from the sky to space for various latitudes [16] compared with the simulated values from a one-dimensional energy balance model.

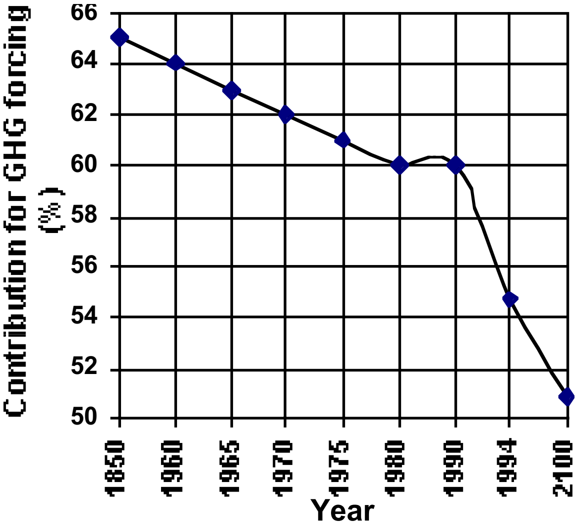

Figure 11
CO_2 contribution to greenhouse has forcing from 1980 to present and projected values to 2100

Again, there is no significant different observed at the 99% confidence level. Modeled values are slightly higher than observed, possibly due to the misrepresentation of the greenhouse effect of clouds by the empirically determined constants used in the equations driving the model.

Radiative Forcing
The global mean temperature change due to different radiative forcing mechanisms have a big range of values due to the uncertainty of the feedback term as well at the size of the forcing.

Figure 7 demonstrates that the combination of GHG forcing over rides other negative forcing because the GHG's have a longer lifetime than aerosols (e.g. N_2O has a life time of about 120 years and a radiative forcing 200 times that of CO_2). Even though the model is rather 'crude', it appears to be moderately reliable.

Calculations of the contribution of CO_2 in GHG forcing is shown in Figure 11. The importance of CO_2 forcing is gradually decreasing, with a sudden decrease from 1960 onwards. Again, the large lifetime and size of forcing, and increase in non-CO_2 GHG, such as N_2O and CH_4, explain this decrease in CO_2 importance. According to Figure 11, by 2100, CO_2 will represent only approximately 50% of the GHG forcing. This means that the non-CO_2 are becoming increasingly important for future climatic changes. The Kyoto Protocol only intends to reduce a 'basket' of six GHG's. Stabilizing CH_4 and N_20 at today's levels produce reductions in anthropogenic emissions of 8% in CH_4 and 50% in N_20 according to our projections. This makes the 5% reduction agreed to in Japan seem to be an irrelevant conquest.

SUMMARY

Although its reliability is still questionable, climate modeling is one of the best tools we have to predict future environmental change. This tool should continue to be investigated to develop mitigation policies designed to minimize the impacts of climate change on human society. The biggest problem in the models is the lack of knowledge about certain climatic processes.

The models created is this project are simple and revealed moderate specificity. We showed that non-CO_2 GHG's are becoming increasingly important in determining future climate changes. Our projection is that the global temperature should increase by around 3 C by 2100. According to projections from our models, the recently agreed upon Kyoto Protocol will diminish this warming by only 0.9 C. These results should be considered with caution since this exercise is oversimplified due to time constraints. We intend to perform a more elaborate exercise in the future.

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