CLIMATE CHANGE

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ABSTRACT

Over the past century global average temperature has increased by about 0.8 °C. Any substantial change in the Earth's temperature must be the result of a perturbation, or so-called radiative forcing, of the planet's energy balance. The equilibrium global temperature response to a particular amount of radiative forcing is termed the Earth's *climate sensitivity*. Once the Earth's energy balance is perturbed, feedbacks arise that act either to enhance or suppress the perturbation. The positive feedback arising from changes in the water vapor level in the atmosphere resulting from a change in temperature is key. In this article, we evaluate the possible causes of the recent warming of the Earth. We review measured variations in the solar output, as well as the historical paleoclimate record, especially the glacial-interglacial cycles. Atmospheric greenhouse gases, carbon dioxide, methane, and nitrous oxide, have increased substantially over the last century; the radiative forcing resulting from these increases can be accurately calculated. The forcing attributable to the increase in levels of airborne particles (aerosols) is considerably more uncertain. All these factors must be taken into account to arrive at an explanation of the recent warming.

CONTENTS

1. INTRODUCTION

Over the last century or so, atmospheric concentrations of so-called greenhouse gases (GHGs), carbon dioxide, methane, nitrous oxide, and halocarbons, have risen dramatically, and global mean temperature has increased by about 0.8°C. That the rise of greenhouse gas concentrations over this time period is a result of human activity is incontrovertible. This article reviews the evidence for the attribution of the global temperature increase. We address the following issues:

1. What is the perturbation to the Earth's energy balance that has occurred as a result of the build-up of GHGs?
2. Water vapor is the dominant greenhouse gas in the Earth's atmosphere; what role does it play in climate change?
3. To what extent are changes in solar irradiance and the galactic cosmic ray flux responsible for the Earth's temperature increase over the past century?
4. What does the paleoclimate record reveal about the response of the Earth's climate to forcings?
5. Aerosols (airborne particles) exert a cooling effect on climate; what role

do they play in the observed temperature change of the planet?

6. And, finally, to what causes can the warming experienced by the Earth over the last century, and the last three decades in particular, be attributed?

2. THE EARTH'S ENERGY BALANCE

Annually averaged, the Earth receives 342 W/m^2 of solar radiation at the top of the atmosphere (Figure 1). Of this amount, 107 W/m^2 is reflected back to space by the surface of the Earth, and by clouds and particles (aerosols) in the atmosphere. As a result, the net radiant energy absorbed by the Earth is 235 W/m^2. At thermal equilibrium, 235 W/m^2 of energy must be radiated

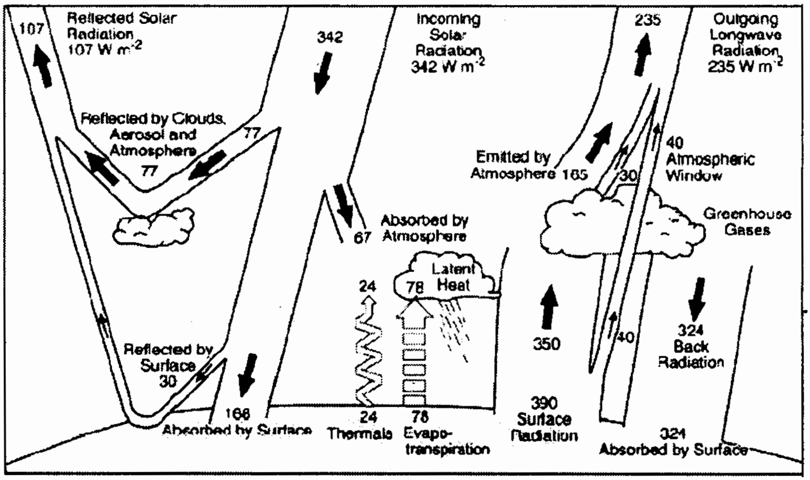

Fig. 1: Earth's energy balance (Kiehl and Trenberth, 1997). Incoming and outgoing energy fluxes from Earth on an annual-average basis. The greenhouse effect refers to the absorption and re-radiation of energy by atmospheric gases, resulting in a downward flux of infrared radiation from the atmosphere to the surface. At equilibrium, the total rate at which energy leaves the Earth (107 W/m^2 of reflected sunlight plus 235 W/m^2 of infrared radiation) is equal to 342 W/m^2 of incident sunlight.

back to space from the Earth, establishing an equilibrium temperature of the Earth.[1] The Earth's surface emits 390 W/m^2 of infrared radiation; much of this is absorbed by gases and cloud droplets in the atmosphere and reradiated back to the Earth's surface; at equilibrium, an infrared radiative flux of 235 W/m^2 escapes from the top of the atmosphere. The Earth's climate is the average state of the atmosphere, land, and water on time scales of seasons and longer.

2.1 Radiative forcing

Radiative forcing, measured in units of watts per square meter (W/m^2), is any imposed perturbation on the Earth's energy balance. It is the imbalance caused by the forcing agent between the solar energy absorbed by the Earth and thermal emission by the Earth back to space. Therefore, radiative forcing is the amount by which the forcing mechanism would change the top-of-the-atmosphere energy budget, if the temperature were not allowed to change so as to restore equilibrium. A forcing is taken as positive if it tends to make the Earth warmer, that is, if the solar input exceeds the thermal (infrared) output. An increase in the luminosity of the Sun is an example of a positive forcing. A large volcanic eruption that injects particles into the lower stratosphere, which reflect more sunlight back to space than in non-volcanic periods, is an example of a negative forcing, which is expressed as negative in sign.

2.2 Climate sensitivity

The global mean temperature change that results in response to an imposed perturbation on the Earth's energy balance after a time sufficiently long for both the atmosphere and the ocean to come to thermal equilibrium is termed the Earth's *climate sensitivity*. Climate sensitivity has units of °C per W/m^2 (°C / W/m^2) and is given the symbol λ.

An estimate of the Earth's climate sensitivity can be obtained from the historical record. The temperature change between full glacial and interglacial conditions was ~ 10 °C in Antarctica, ~ 3 °C in the Pacific warm

[1] The equilibrium climate is, more precisely, a quasi-equilibrium climate, owing to the fact that changes do occur on millennial and longer timescales, such as glacial-interglacial cycles caused by orbital variations.

pool, and ~ 5 °C global average. The estimated difference in forcing between the two periods[2] was ~ 6.6 W/m^2. This gives a climate sensitivity of

$$\lambda = \frac{5^\circ C}{6.6\ W/m^2} \sim 0.75 \frac{^\circ C}{W/m^2}$$

As we will see shortly, the atmospheric concentration[3] of CO_2 has increased from its pre-industrial level of 280 parts-per-million (ppm) to a current level of 380 ppm. A standard benchmark that is used to assess climate change is a doubling of CO_2 from its pre-industrial level to 560 ppm. (This scenario is referred to in shorthand notation as 2xCO_2.) The perturbation to the outgoing infrared radiation from the Earth that would result from 2xCO_2 is 3.7 W/m^2. The increase in the global mean temperature required to re-equilibrate the Earth's energy balance to this change, considering solely the Earth's blackbody Stefan-Boltzmann response, would be 1.2 °C. A temperature increase of 1.2 °C in response to a forcing of 3.7 W/m^2 implies a climate sensitivity of 0.32 °C/W/m^2. This value can be compared to the climate sensitivity estimated above for the glacial-interglacial cycles of ~ 0.75 °C/W/m^2. The temperature response of the Earth to a 2xCO_2 forcing, as predicted by a climate sensitivity of 0.75 °C / W/m^2, is about 2.8 °C. The estimated global mean temperature increase of 2.8 °C from a doubling of CO_2 represents an amplification by a factor of about 2.3 over the purely radiative effect of 1.2 °C. What is the reason for this discrepancy?

The explanation of why the actual temperature change is so much larger than that based purely on the amount of absorbed radiation lies in *climate*

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[2] At the peak of the ice ace, it is estimated that the global forcing as compared to non-glacial times was −6.6 W/m^2. This amount of cooling comprised −3.5 W/m^2 due to ice sheets, −2.6 W/m^2 due to lower greenhouse gas levels, and −0.5 W/m^2 due to increased levels of aerosols (dust).

[3] It is conventional to describe atmospheric concentrations in terms of dimensionless mixing ratios, such as parts-per-million (ppm). A mixing ratio is just the mole fraction, or, equivalently, the volume of substance per unit volume of air. A mixing ratio of 280x10^{-6} is written as 280 ppm. Concentration should be expressed as, for example, molecules/cm^3; nevertheless, "mixing ratios" tend to be used interchangeably with "concentrations".

feedbacks. Feedbacks are effects that arise as a result of the forcing. A climate forcing that causes warming, for example, may lead to a melting of some of the sea ice. Because the darker ocean absorbs more sunlight than the sea ice it replaced, this leads to further warming; the result is a positive feedback. In a real sense, the single most important concept in understanding climate change is that of feedbacks. Much of this review is devoted to explaining climate feedbacks and exploring their role in past and present climate.

Once a constant forcing is imposed on the Earth's radiative energy balance, the Earth's climate will adjust to a new temperature equilibrium. The thermal response times of the climate components are:

Land	~ 1 week
Atmosphere	~ 1 month
Ocean surface layer (top 100 m)	~ 10 years
Deep ocean	~ 1000 years

The thermal heat capacity of the ocean surface layer vastly exceeds that of the land and the atmosphere. For practical purposes, the response time to a perturbation in the Earth's energy balance depends essentially on the time needed for the ocean surface temperature to reach equilibrium. With a characteristic response time of ~10 years, several decades or so are required for the ocean surface layer to reach thermal equilibrium with the atmosphere. Over a millennial time scale, the surface ocean and deep ocean approach thermal equilibrium.

If the forcing is changing on a time scale shorter than a few decades or so (which is the case with CO_2), the climate does not have a chance to "catch up" and equilibrate. As a result, the temperatures in the atmosphere and surface ocean will not be at equilibrium. The atmosphere cannot warm completely until the surface ocean warms, which requires several decades. Thus, the actual temperature increase (in the case of a positive forcing) lags behind the ultimate equilibrium temperature increase. In the terminology of climate change, the *transient* temperature lags behind the *equilibrium* temperature.

3. GREENHOUSE GASES AND OTHER SUBSTANCES

Many atmospheric gases have the ability to absorb infrared radiation, but only those present at significant concentrations can affect appreciably the Earth's energy balance. Water vapor is, in fact, the major atmospheric absorber of infrared radiation. The most important greenhouse gases in the Earth's atmosphere, besides H_2O, are carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), ozone (O_3), and the halocarbons, the so-called well-mixed GHGs.

The absorption spectra of atmospheric gases are instrumental in the chemistry of the atmosphere and the Earth's radiation balance. Oxygen and ozone absorb radiation strongly in the ultraviolet region, but their absorption is essentially zero in the visible and infrared regions. Methane absorbs strongly in two narrow regions around 3.5 and 8 μm wavelength, which are in the infrared portion of the spectrum. Nitrous oxide has absorption peaks at about 5 and 8 μm. Carbon dioxide has a more complex absorption spectrum with isolated peaks at about 2.6 and 4 μm and a broad shoulder above about 13 μm. The absorption spectrum for water vapor is even more complex than that of CO_2, with numerous broad peaks in the infrared region between 0.8 and 10 μm. (Solar radiation can be approximated as that from a blackbody at 5780 K, and that from the Earth as emitted by a blackbody at 255 K. Infrared radiation emitted by the Earth has a maximum near 10 μm.) Water vapor is so abundant in the atmosphere that in those regions of the spectrum where H_2O vapor absorbs infrared radiation, the spectrum is saturated. CO_2 and CH_4 have absorption in some of the "windows" in the infrared spectrum where H_2O vapor does not absorb and where terrestrial radiation escapes. At the year 2003 level of CO_2 (375 ppm), absorption by CO_2 is not saturated in its frequency bands, so that additional CO2 will lead to additional absorption of infrared radiation (Figure 2).

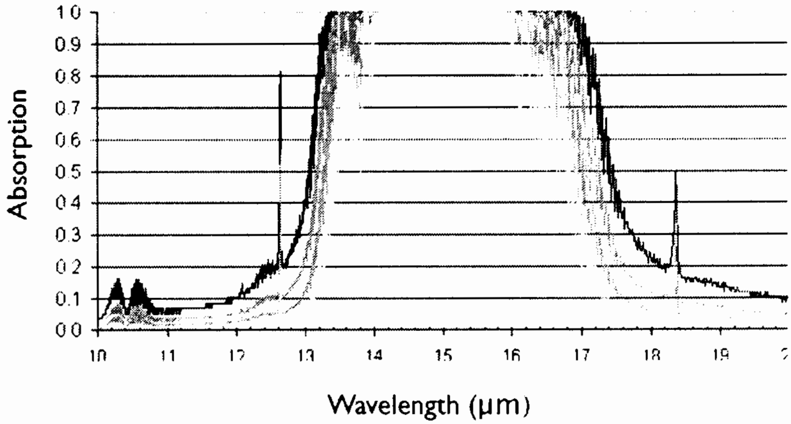

Fig. 2: Absorption by CO_2 in the infrared spectral range of 10 to 20 μm. The four curves, starting from the lowest, represent atmospheric CO_2 levels that are a factor of 0.5, 1.0, 2.0, and 4.0 of 375 ppm. Between 14 and 16 μm, the spectrum is saturated. For wavelengths less than 14 μm and greater than 16 μm, additional CO_2 leads to additional absorption.

The mechanism by which the Earth loses energy is infrared radiation to space. The troposphere is characterized by convection that constantly lifts air from the surface to higher altitudes; in the process of rising, the air cools by expansion. As a result, the temperature decreases with height in the troposphere. Convection and other heat transport mechanisms connect all levels in the troposphere, and, to a good approximation, the troposphere can be considered to warm and cool as a unit. When a greenhouse gas is present, the atmosphere will be partly opaque to infrared radiation, and the upwelling infrared radiation is absorbed before it has a chance to get very far above the surface. As a result, the infrared radiation that actually escapes to space comes from the higher, colder parts of the atmosphere. Since the emission rate of radiation from a blackbody varies with the fourth power of temperature, the flux of radiation from these upper levels is considerably less than that which is emitted from the surface. By contrast, the downwelling radiation to the surface comes predominantly from the warmer layers nearest the surface.

Because the radiation to space and that to the surface originate from different levels in the troposphere, an increase in GHG concentration leads to two different effects. As noted, as GHG levels increase, the atmosphere becomes more opaque to infrared, and the infrared escapes from higher and colder parts of the atmosphere, leading to less outgoing radiation. The result is that the planet receives more solar energy than it loses. Solar energy is primarily absorbed at the Earth's surface and transferred to the atmosphere by surface heat fluxes. In order to achieve equilibrium, the troposphere must warm until the overall energy budget is restored. Since the entire troposphere warms and cools together, this means that the air near the surface must warm as well.

The change in air temperature near the surface then translates into a change of the temperature of the surface itself through three mechanisms: (1) sensible heat flux; (2) latent heat flux (cooling the surface by evaporation); and (3) infrared heat flux (cooling by emission and warming by absorption). The latent heat flux tends to be the dominant term. In warm, wet areas, evaporative heat transfer is so effective that the surface temperature remains close to the overlying air temperature, and, in effect, the surface temperature follows that of the troposphere just above the surface. By contrast, the daytime surface of the Sahara tends to be about 10° C warmer than the overlying air, because in the absence of moisture, the relatively inefficient sensible and radiative heat transfer require a relatively large temperature difference to generate the necessary heat flux. For this reason, if relatively moist mid-latitude areas dry out in response to increased warming, the reduction in soil moisture will compound the surface temperature increase. In summary, the top of the atmosphere energy budget governs the warming of the low level air temperature, whereas the surface energy balance determines the difference between the air temperature and the surface temperature.

Figure 3 shows atmospheric concentrations of CO_2, CH_4, and N_2O over the last 2000 years. CO_2 has increased from fossil fuel use and deforestation, the latter of which both releases CO_2 and reduces its uptake by plants. The CH_4 increase is a result of agricultural activities, natural gas distribution, and landfills, as well as natural processes in wetlands. N_2O is emitted from fertilizer use and fossil fuel burning, as well as from natural processes in soils. The fourth class of long-lived GHGs is halocarbons, which were used extensively as refrigeration agents and in other industrial applications before their role in stratospheric ozone depletion was discovered. Their concentrations are now decreasing as a result of international regulations to limit their use.

Global atmospheric concentrations of three well mixed greenhouse gases

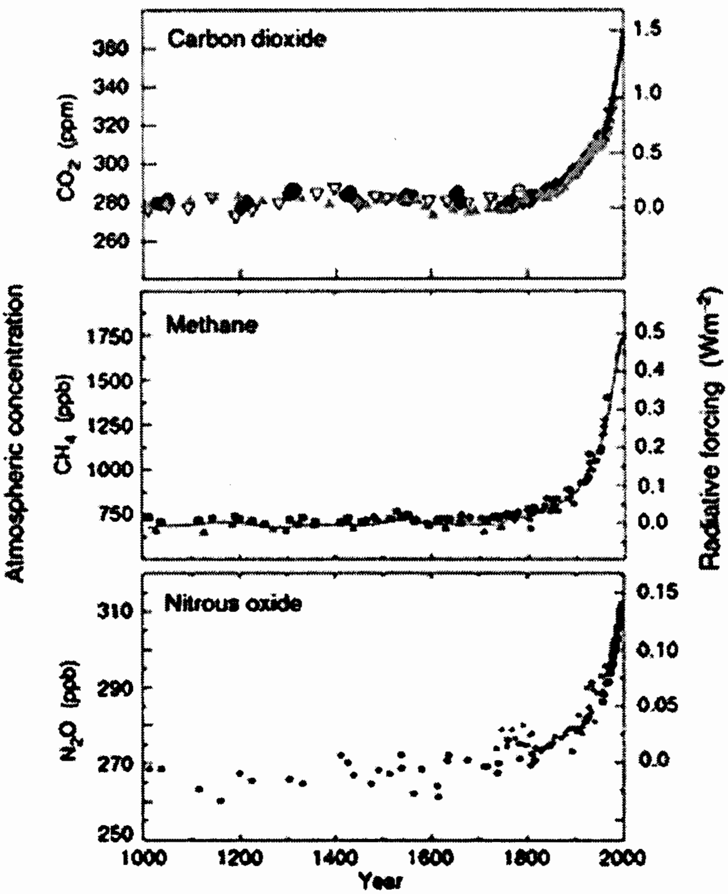

Fig. 3: Atmospheric concentrations and radiative forcing of carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O) over the last 1000 years.

That the increase in the atmospheric level of CO_2 over the past century or so is a result largely of fossil-fuel burning is conclusively demonstrated from several lines of evidence. These include: (1) records of coal, oil, and natural gas consumption; and (2) concomitant decreases in the relative abundance of both the stable (^{13}C) and radioactive (^{14}C) carbon isotopes and the decrease in atmospheric oxygen. If heating of the oceans were the source of atmospheric CO_2, a trend of decreasing concentration of CO_2 would be expected; yet, what is observed is the opposite - CO_2 concentrations in the ocean have risen even as ocean temperature has risen (Tedesco et al., 2005).

Carbon in CO_2 has two naturally occurring stable isotopes, ^{12}C and ^{13}C. ^{12}C constitutes about 99% of the C in CO_2, with ^{13}C being about 1%. The $^{13}C/^{12}C$ ratio in CO_2 emitted from fossil fuel combustion is less than that in atmospheric CO_2, so that when CO_2 from fossil fuel combustion enters the atmosphere, the $^{13}C/^{12}C$ ratio in atmospheric CO_2 decreases at a rate that can be predicted based on the magnitude of fossil fuel emissions. (The $^{13}C/^{12}C$ ratio in CO_2 can be measured at 1 part in 10^5.) The energetics of photosynthesis lead to a preference for $^{12}CO_2$ over $^{13}CO_2$, as slightly less energy is required to bond to $^{12}CO_2$ than $^{13}CO_2$. Thus, the naturally occurring $^{13}C/^{12}C$ is skewed toward ^{12}C in plants. Fossil fuels were originally plants, which explains why fossil fuel CO_2 is depleted in ^{13}C. ^{13}C fractions have also decreased in ocean surface waters over the past decades. Such behavior is consistent with an atmospheric fossil fuel source of CO_2 and inconsistent with an oceanic source for the following reason. Oceanic carbon has slightly more ^{13}C than atmospheric carbon, but $^{13}CO_2$ is heavier and less volatile than $^{12}CO_2$; as a result, CO_2 degassed from the ocean has a ^{13}C fraction close to that of atmospheric CO_2. It is not possible therefore, that an oceanic source of CO_2 could lead to a simultaneous drop of ^{13}C in both the atmosphere and the ocean.

The unstable carbon isotope, ^{14}C, termed ***radiocarbon***, comprises about 1 in 10^{12} carbon atoms in the atmosphere. ^{14}C has a half-life of 5700 years. The stock of ^{14}C is replenished in the upper atmosphere by the interaction of cosmic rays with ^{14}N. Fossil fuels contain no ^{14}C, as it decayed long ago. Emission of CO_2 from burning fossil fuels thus lowers the atmospheric ^{14}C fraction. Atmospheric ^{14}C, measured in the tree rings, decreased by 2 to 2.5% from about 1850 to 1954, when atmospheric nuclear testing began to inject ^{14}C into the atmosphere. The observed decline in ^{14}C cannot be explained by a CO_2 source from terrestrial vegetation or soils.

CO_2 is not destroyed chemically, and its removal from the atmosphere occurs by uptake in the land and ocean reservoirs, and ultimately as mineral deposits. The lifetime of CO_2 in the atmosphere is governed by uptake to these surface reservoirs but over a millennial time scale is controlled by transfer from near-surface waters to the deep ocean.

Both CO_2 and CH_4 were trapped long ago in air bubbles preserved in Greenland and Antarctic ice sheets, which have survived from the series of ice ages over hundreds of thousands of years. CO_2 concentrations ranged between 190 ppm during the ice ages to about 280 ppm during the warmer interglacial periods, the last of which began about 10,000 years ago. Lowest values of methane concentrations were 0.3 ppm in the depths of ice ages to 0.7 ppm in the warmest interglacials.

Over the course of the Earth's existence, 4.5 billion years, volcanoes and mantle outgassing have released much more CO_2 than that released from the burning of fossil fuels. Annual CO_2 emissions at present, however, from volcanoes and human activities, in metric tons (Mt), are roughly:

Volcanoes	$\sim 1.0 \times 10^8$ Mt / yr
Humans	$\sim 1.0 \times 10^{10}$ Mt / yr

The present-day volcanic source of CO_2 thus represents only about 1% of that attributable to humans.[4]

The strength of climate forcing for a greenhouse gas can be expressed as W/m^2 of forcing per unit of increase of atmospheric concentration. The relative strengths of CO_2, CH_4, and N_2O at their present levels, expressed as W/m^2 per ppm, are:

CO_2 (380 ppm)	0.0044	W/m^2 / ppm
CH_4 (1.8 ppm)	0.2706	
N_2O (320 ppb)	0.5106	

Per part-per-million increase, CH_4 and N_2O are 62 and 114 times as large as CO_2; however, CH_4 is 200 times less abundant than CO_2, and N_2O is $\sim$ 1000 times less abundant, so CO_2 exerts the dominant effect.

[4] Some have claimed that atmospheric CO_2 increases over the past century are the result of volcanic emissions, not anthropogenic activity. Were this the case, the massive eruption of Mt. Pinatubo in 1991 would have been accompanied by a large pulse in CO_2 and associated longer-term warming. No CO_2 pulse appeared in any global measurements, and, in fact, the main effect of the eruption was a temporary cooling from the resulting stratospheric aerosols.

3.1 Global forcings of GHGs and other substances

Table 1 gives estimated present-day global forcings from GHGs and other substances. CO_2 is the largest single contribution with a forcing of +1.66 W/m^2. Other GHGs, methane, halocarbons, and nitrous oxide, together, exert a climate forcing of +0.97 W/m^2. Methane is oxidized in the atmosphere by hydroxyl radicals to CO, and eventually to CO_2, with an atmospheric lifetime of about 10 years. Increases in CH_4 lead to an increase in stratospheric water vapor (about 7% of CH_4 is oxidized in the upper atmosphere) and to an increase in tropospheric O_3 through reactions involving oxides of nitrogen, thus having an indirect effect on forcing through stratospheric water vapor and ozone as well.

Table 1

Radiative forcings since pre-industrial time (IPCC, 2007)

Species	Radiative forcing (W/m^2)
Long-lived greenhouse gases	
CO_2	+1.66 ± 0.17
CH_4	+0.48 ± 0.05
N_2O	+0.16 ± 0.02
Halocarbons	+0.34 ± 0.03
Total	+2.63 ± 0.26
Ozone	
Stratospheric[a]	−0.05 ± 0.10
Tropospheric	+0.35 [−1.0, + 0.3]
Stratospheric H_2O vapor from CH_4	+0.07 ± 0.05
Aerosol	
Total direct	−0.50 ± 0.40
Indirect (cloud albedo)	−0.70 [−1.1, + 0.4]
Solar irradiance	+0.12 [−0.06 + 0.18]

[a] A reduction in stratospheric O_3 affects surface temperature in two ways: (1) less stratospheric O_3 implies that more solar radiation will reach the surface-lower atmosphere system, which will tend to warm the climate; but (2) less stratospheric O_3 will lead to a cooler stratosphere (due to less absorption of solar radiation by O_3) that leads to less downwelling infrared radiation to the surface-lower atmosphere system, which will tend to cool the climate. The net effect on climate is smaller than either of the two effects taken alone, and the sign of the net effect depends on the altitudes where the O_3 change takes place. A reduction in O_3 at higher altitudes leads to net warming; a reduction of O_3 at lower altitudes leads to net cooling.

Atmospheric aerosols play an important role in the global climate system through modifications of the global radiation budget: (1) directly, mainly by scattering and absorption of solar radiation, as only substantial concentrations of large aerosols at higher altitudes cause a significant long-wave direct radiative effect, and (2) indirectly, by the modification of cloud properties and abundance. In particular, the interactions between aerosols and clouds introduce considerable uncertainty to estimates of the overall aerosol radiative effects.

The term "indirect aerosol effect" is generally used for mechanisms by which aerosols affect the cloud microphysical properties via the availability of cloud condensation nuclei (CCN), depending on their size-distribution, composition, and mixing state. The most prominent indirect aerosol effects are: (1) the cloud albedo effect, i.e. the effect of aerosols on the cloud droplet number concentration, thus, under the assumption of a fixed liquid water path, on the droplet size and consequently cloud albedo; (2) the cloud lifetime effect. The cloud lifetime effect is that in which increased cloud droplet number concentrations and smaller droplets, owing to increases in aerosol concentration, lead to a suppression of formation of collisionally-induced precipitation-sized droplets, leading to a longer cloud lifetime. The cloud lifetime effect is strictly not an external perturbation but rather a first-order feedback process. At present, clear observational support for the cloud lifetime effect is lacking, and the possible level of global radiative forcing due to it cannot be estimated at this time.

In addition to the indirect aerosol effects, the aerosol direct radiative effects cause first-order responses of the hydrological cycle by two key mechanisms: (1) reducing the surface insolation, thus altering the surface fluxes, as well as (2) by heating the atmosphere, thus altering profiles of temperature and relative humidity and consequently the atmospheric stability. While only absorbing aerosols, i.e. aerosols with sufficiently low single-scattering albedo, directly heat the atmosphere, both absorbing and exclusively scattering aerosols reduce the surface insolation. Finally, the spatially and temporally highly inhomogeneous direct and indirect aerosol radiative effects cause changes in the large-scale dynamics that, in turn, affect the hydrological cycle and cloud distribution, a higher order response of the climate system.

One might ask - What is the magnitude of the direct energy release by human energy production and how does it compare with GHG forcing? As noted above, the average annual energy input to the Earth from the Sun is about 235 W/m^2. The amount of energy released annually as a result of humanity's energy production is about 0.025 W/m^2 (Nakicenovic et al., 1998). This can also be compared with the internal terrestrial energy generation of 0.087 W/m^2 (Pollack et al., 1993). When the human energy production is compared with the anthropogenic GHG forcing of ~ 3 W/m^2, it is noted that global anthropogenic heat release is totally negligible in the Earth's energy balance.

3.2 The role of water vapor in climate change

Water vapor is, of course, a natural constituent of the Earth's atmosphere, and it is the principal absorber of long-wave infrared radiation emitted from the surface of the Earth. Even at the current level of atmosphere CO_2, water vapor makes up between 60% and 70% of the greenhouse effect; CO_2 constitutes between 20% and 30%. The amount of water vapor in the atmosphere is determined by a balance between evaporation, principally from the oceans, and removal via precipitation. The mean atmospheric lifetime of a H_2O molecule is about 10 days; this means that, following any perturbation, a new water vapor equilibrium is established in a couple of weeks. (Too much water vapor in the air will rain out; too little, and more evaporates from the oceans, over this time scale.) If the global mean temperature of the Earth were to change, the water vapor level in the atmosphere adjusts to maintain a constant relative humidity. (That the water vapor concentration over a deep region of the troposphere responds to temperature changes in such a way as to maintain a constant relative humidity is not a trivial result and is beyond our scope here (See Pierrehumbert et al., 2007).) Say that global temperature increases as a result of a positive climate forcing (e.g. CO_2 increase or increase in solar irradiance). At the higher temperature, a constant relative humidity translates to a higher absolute concentration (specific humidity), so the total amount of water vapor in the air increases. Specifically, the water holding capacity of the atmosphere increases by 7% for each 1 °C rise in temperature (as predicted by the Clausius-Clapeyron relation).[5] And once the

[5] For water vapor at terrestrial temperatures, the Clausius-Clapeyron relation is well approximated by

air is warmed by other means, H_2O concentrations rise and stay high.

Because H_2O vapor is the dominant greenhouse gas, once its concentration rises as a result of an external forcing that raises the Earth's temperature, the additional infrared absorption by the increased H_2O amplifies the initial warming through a powerful feedback effect (See Appendix). It is the feedback effects, of which water vapor is the dominant one, that explain why the temperature change due to a doubling of CO_2 is about 2.5 times that attributable to infrared absorption by CO_2 alone. A test of climate models' ability to simulate the water vapor feedback effect was afforded after the 1991 eruption of Mt. Pinatubo in the Philippines (Soden et al., 2002). The volcanic aerosol caused global cooling for more than two years after its eruption. From the observed amount of sulfate aerosol emitted by the volcano, models were able to predict the observed decrease in water vapor resulting from the cooling induced by the aerosol. In summary, the climate role of water vapor is a classic positive feedback effect, which would not occur in the absence of forcing from some other agent.

Climate models predict that the concentration of H_2O vapor will increase in the upper troposphere as a result of increases in GHGs. Satellite measurements show a signature of upper tropospheric moistening over the period 1982 to 2004 (Soden et al., 2005), in accordance with predictions. This result is especially important in establishing the validity of treatment of water vapor feedback by climate models. Water vapor feedback is the result of water vapor changes on the top-of-the-atmosphere radiation budget. Water vapor near the surface has little influence on the top-of-the-atmosphere balance, because the temperature of the air near the surface is close to that of the surface itself. By contrast, the relatively small quantity of water vapor aloft has a substantial influence on the top-of-the-atmosphere energy budget, because it increases the infrared opaqueness of these layers that are much colder than the surface. In so doing, this H_2O vapor blocks the upwelling infrared from the warmer parts of the atmosphere and replaces it with

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$$p_s(T) = p_s(T_0)\exp\left[-\frac{L}{R}\left(\frac{1}{T}-\frac{1}{T_0}\right)\right]$$

where $p_s(T)$ is the saturation vapor pressure of water at temperature T, L is the latent heat of vaporization, R is the gas constant, and T_0 is a reference temperature (L/R = 5419 K). At the freezing point, p_s is 614 Pa; at 300 K, p_s rises to 3664 Pa.

infrared emissions from the cold layer. Soden et al. (2005) analyzed satellite data to compare mid-to upper tropospheric H_2O vapor observations with general circulation model predictions for the 1982 - 2004 period. The data indicate that upper-level moisture increases in warmer conditions, in the same manner as predicted by models. Moreover, by artificially suppressing moisture changes in synthetic data, they were able to definitively reject the hypothesis that upper troposphere water vapor content would remain constant as temperature increases. (In an attempt to dismiss the link between GHGs and global warming, some have suggested that upper tropospheric water should remain constant as the atmosphere warms, in contrast to the behavior predicted by climate models; the work of Soden at al. (2005) decisively negates this argument.)

Finally, it is important to point out that the radiative effects of water vapor are predicted by basic radiation physics, such as Kirchoff's Law and band models of radiation, which are well confirmed in the laboratory, by observations, and by comparison with line-by-line radiative calculations. The essence of the radiative effect of the water vapor distribution is that water vapor at a high, cold level blocks the infrared redistribution from below and replaces it with infrared emitted at a lower temperature. Adding more water vapor near the surface, where the air temperature is nearly the same as that of the surface, does not have this effect since each radiating surface is at essentially the same temperature. The radiative effects of water vapor are approximately logarithmic in concentration (like most GHGs), so a doubling or halving of the relatively small amount of water vapor in the mid to upper troposphere is radiatively significant.

Clouds affect the radiative energy balance of the planet. Cloud radiative forcing is just the difference between the clear-sky and cloudy-sky energy gains. The solar and terrestrial radiative properties of clouds have offsetting effects in terms of Earth's energy balance. In the infrared, clouds generally reduce the radiation emission to space and thus lead to a heating of the planet. In the solar region of the spectrum, clouds reflect sunlight back to space and thus produce a cooling. In areas covered by deep cumulus clouds, almost all of the incoming solar radiation is reflected back to space. The cold cloud tops radiate very little energy to space. High, thin cirrus clouds, which often form from the outflow of moist air in the cumulus anvils, reflect some solar radiation back to space but also let some through to the Earth's surface. They emit some longwave energy both to space and back to the surface.

The Earth Radiation Budget Experiment (ERBE), employing the NASA Earth Radiation Budget Satellite, launched in 1984 (Barkstrom et al., 1989), has provided key data on the net radiative cloud effect. Results from ERBE indicate that, at present in the global mean, clouds reduce the radiative heating of the planet. The degree of cooling depends on season and ranges from about −13 to − 21 W/m^2. On the basis of hemispheric averages, longwave and shortwave cloud forcings tend to balance each other in the winter hemisphere. In the summer hemisphere, the negative shortwave cloud forcing dominates the positive longwave cloud forcing, leading to a net cooling.

A hypothesis concerning water vapor and cloud feedback in the tropics was advanced by Lindzen et al. (2001). Focusing on the Indo-Pacific Warm Pool, the portion of the Pacific Ocean from 30°N to 30°S latitude and 130°E to 170°W longitude, from satellite data they concluded that the amount of high thin cirrus clouds varies systematically with changes in sea surface temperature, in such a manner that as sea surface temperature increases, the area covered by high cirrus clouds contracts. Lindzen's group likened this response to an adaptive infrared iris: much like the iris in a human eye contracts to permit less light into the pupil in a brightly lit environment, cirrus clouds contract to allow more infrared heat to escape as temperature increases. This would constitute a negative feedback, since the cloud changes would act to retard a temperature increase and stabilize the climate system. Lindzen et al. argued that the magnitude of the infrared iris effect could cancel the positive feedback of water vapor, which would reduce climate sensitivity to that attributable solely to the Earth's infrared blackbody response.

Scientists immediately sought to confirm Lindzen's iris hypothesis using data from the Clouds and the Earth's Radiant Energy System (CERES) sensor, on board the NASA Tropical Rainfall Measuring Mission (TRMM) satellite, launched in 1997 and subsequently on satellites launched in 1999 and 2002. CERES is the most advanced space-based sensor for measuring Earth's radiant energy fluxes. The analysis of CERES data by Lin et al. (2002) and Chambers et al. (2002) produced results counter to those presented by Lindzen et al. (2001). These investigators found that, while tropical clouds do change in response to a warmer sea surface temperature, they do so in such a way as to slightly enhance warming at the surface. Lindzen et al. predicted that an extra 70 W/m^2 of heat is emitted to space from the Earth per unit tropical, cloudy area that disappears with increased sea surface temperature; the CERES data indicate a +2 W/m^2 warming. The CERES data revealed that the cloud albedo in this region is about 51%, rather than 35% as assumed in the Iris calculation; the result of this is that the infrared greenhouse effect exerted by the clouds is less than that assumed in the Iris calculation. Another disagreement between the studies of Lindzen et al. (2001) and Lin et al. (2002)/Chambers et al. (2002) pertains to the amount of heat escaping to space from cloudy regions. The CERES data show that, on average, 155 W/m^2 of infrared energy leaves the atmosphere over cloudy, moist regions versus 138 W/m^2 as assumed in the Lindzen et al. paper. Fu et al. (2002) also evaluated the Iris model, in which they argue that water vapor feedback is overestimated by at least 60%, and that the high cloud feedback is small. Although some continue to endorse the Iris Hypothesis, the weight of evidence in the literature and the lack of agreement with CERES data suggest that the original hypothesis was substantially overestimated.

4. SOLAR OUTPUT

Understanding the relationship between solar variability and the Earth's climate is of prime importance when assessing the anthropogenic role in climate change (Benestad, 2002; Meehl et al., 2003). Solar irradiance varies slightly over an 11-year cycle due to variation in the Sun's magnetic activity. Sunspots, dark regions on the solar disk, have been used to track fluctuations in the strength of the Sun's 11-year activity cycle for almost 400 years. Variations in satellite records of lower troposphere temperatures since 1978, in upper ocean temperatures since 1955, and in surface temperatures during the past century are approximately in phase with this 11-year solar cycle (Lean and Rind, 2001). More sunspots equate to higher solar output. Figure 4 shows sunspot observations over the last 400 years. The so-called Maunder Minimum, or Little Ice Age, which lasted from 1650 to about 1750, was characterized by very low sunspot activity and unusually cold temperatures over Europe.

Since 1978, solar irradiance has been measured with high precision from satellites; prior to that, sunspot observations provide a measure of solar activity. In addition, cosmogenic radionuclides, ^{10}Be and ^{14}C, serve as proxies to extend solar activity reconstructions beyond the period of direct observations. Produced in the atmosphere by the interaction of galactic cosmic rays with the atmosphere, these radionuclides have decreased production rates during periods of high solar activity. Their records exist in ice cores (^{10}Be) and tree rings (^{14}C) (Bard et al., 2000).

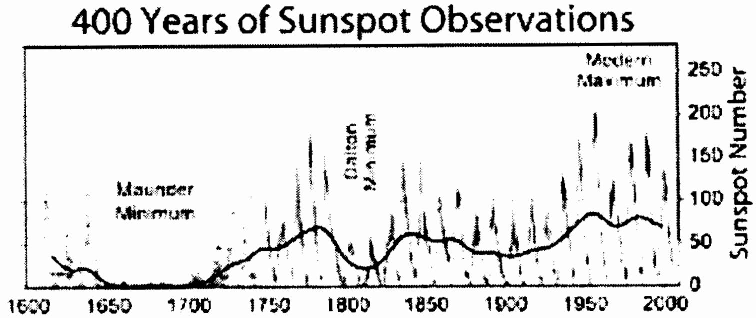
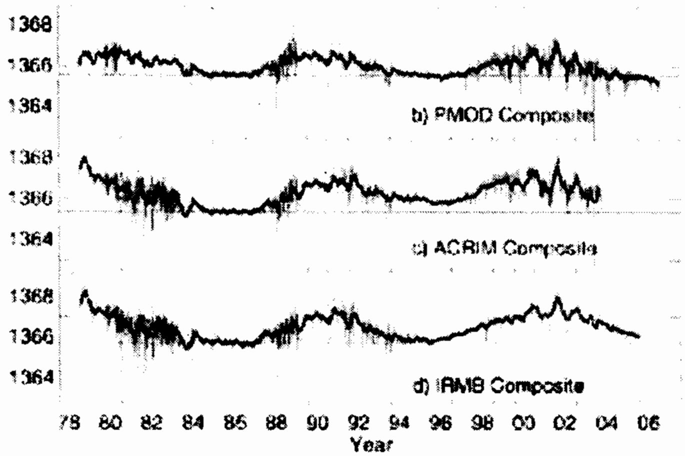


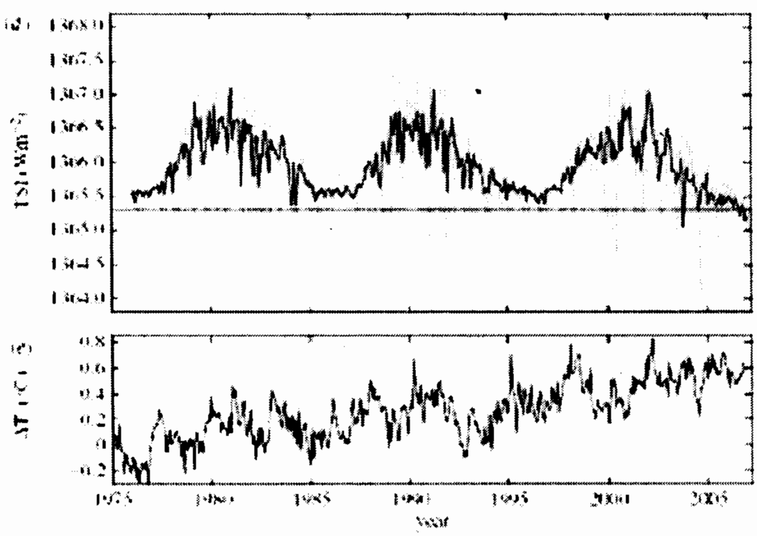
Fig. 4: Record of sunspot observations over the period 1600 - present day. Two periods of especially cool temperatures, the Maunder Minimum (also referred to as the Little Ice Age) ~ 1650 - 1750, and Dalton Minimum ~ 1800 - 1850, correspond to low sunspot number.

4.1 Solar irradiance

The period of greatest warming over the past century or so has occurred since 1970, and if the Sun is responsible for this warming, a sustained increase in solar irradiance would have to be evident in the satellite record. That record (Figure 5) indicates no net increase in solar irradiance since 1978, and reconstructions of pre-satellite data show this period of quiescence extends back to 1940. In a rigorous statistical analysis of solar output and Earth's temperature variations, Krikova and Solanki (2004) assumed that the Sun was totally responsible for the change in Earth's temperature prior to 1970 and that this interplay persisted after 1970. Then, using reconstructions and measured records of solar output, they estimated the fraction of the dramatic temperature rise after 1970 that could be due to the influence of the Sun. In so doing, the solar contribution to global warming prior to 1970 cannot be underestimated, and, as a result, the solar contribution derived for the period after 1970 constitutes an upper limit. The analysis shows that since 1970 the solar influence on climate was not a significant cause of the observed temperature increase.



Upper panel



Lower panel

Fig. 5: Total solar irradiance observations since 1978. Since November 1978, a set of total solar irradiance (TSI) measurements from space is available. From measurements made by several different instruments a composite record of TSI can be constructed. At present, three composites are available, called PMOD, ACRIM, and IRMB. Slight differences in data from different instruments arise

from different calibration procedures and rates of degradation. Description of the procedures used to construct composites can be found in Fröhlich and Lean (1998) and Fröhlich (2000). The upper panel shows three composites: PMOD (World Radiation Center, Davos, Switzerland), ACRIM (Active Cavity Radiometer Irradiance Monitor), and IRMB (Royal Meteorogical Institute of Belgium.) The lower panel shows the composite TSI as daily values, as presented by the World Radiation Center, as compared with the global mean temperature record.

The average surface global variation of temperature over the 11-year solar cycle is ~ 0.2 °C, corresponding to a change in radiative forcing from solar-min to solar-max of ~ 0.18 W/m^2. This amount of temperature change is too large to be explainable by the direct radiative effect of a 0.18 W/m^2 variation in total solar intensity. The answer lies in the positive feedbacks inherent in the climate system, which we have discussed above. The inferred climate sensitivity based on the climate response to sunspot cycles is thus

$$\lambda = \frac{0.2^\circ \mathrm{C}}{0.18\,\mathrm{W/m^2}} \sim 1 \frac{^\circ \mathrm{C}}{\mathrm{W/m^2}}$$

(Analyses of polar warming over sunspot cycles suggests that sea ice-albedo feedback is largely responsible for the amplification that occurs in the polar region.)

A rigorous statistical analysis of solar cycle variations and surface temperature has recently been carried out (Camp and Tung, 2007). With the analysis of Camp and Tung (2007), one can now evaluate the ability of general circulation models to simulate these changes, providing an important test of the validity of general circulation models.

The implications of the derived climate sensitivity of ~ 1 °C / W/m^2 based on the solar cycle are important. Since this climate sensitivity is a *transient*, rather than an *equilibrium* sensitivity (i.e. the climate has not come to equilibrium in response to the varying solar output), and since the equilibrium climate sensitivity always exceeds the transient sensitivity, this implies that the equilibrium climate sensitivity exceeds 1 °C / W/m^2. Thus, for 2xCO_2, with a forcing of ~ 3.7 W/m^2, the equilibrium warming would exceed 3.7 °C. The "high end" predictions for a doubling of CO_2 are usually cited as 4.5 °C; this means that the climate response to solar forcing over the

11-year sunspot cycles implies a climate sensitivity that is near the upper end of those predicted by climate models.

4.2 Cosmic rays

It has been suggested that a connection exists between solar activity, through its effect on galactic cosmic rays, and global warming of the past several decades (Friis-Christensen and Lassen, 1991). The mechanism involves the effect of the cosmic ray flux on the generation of atmospheric ions, which can serve as nucleation sites for formation of ultrafine particles (Svensmark, 1998), some of which then grow large enough to become cloud condensation nuclei (CCN). Thus, the connection involves the relationship between the galactic cosmic ray flux and global cloudiness. Variations in the solar wind modulate the cosmic ray flux; when sunspot number is at its minimum, the cosmic ray flux is at its maximum and vice versa. It is argued that recent global warming is a result of reduced cloud cover owing to variation of the solar-induced cosmic ray flux.

Cosmic rays provide a source of ions in the atmosphere. Above the continental boundary layer, cosmic rays are the principal source of small ions, and cosmic rays are the main source of ions at the surface in marine air. The ionization rate increases towards the poles and with increasing altitude. It is well established that ultrafine particles, so-called condensation nuclei, can be formed from ions. The efficiency of the nucleation of material on atmospheric ions depends on the presence of trace condensable species and the amount of pre-existing particles that compete with the ions for vapor. Ion-induced ultrafine particles can grow by accretion of vapor molecules to reach a size (~ 50 nm diameter or so) at which they are sufficiently large to act as cloud condensation nuclei. The fraction of ultrafine particles that ultimately grows to CCN size depends on the availability of condensable vapors but is generally small. In addition, pre-existing particles may become charged through collision with atmospheric ions. Some evidence suggests that charged aerosols may be more efficient as ice nuclei.

For a demonstrable effect on climate, it would be necessary that the galactic cosmic ray flux exhibit a clear and significant long-term downward trend. Such a trend, however, does not exist (Figure 6). The lack of such a trend does not invalidate the mechanism proposed by Svensmark (1998) that connects the cosmic ray flux to CCN formation, only the proposal that a trend in cosmic rays can explain recent warming. It should be noted, however, that even if fewer CCN were generated by cosmic rays, this does not necessarily mean that global cloud formation would be inhibited, since in many regions of the Earth ample CCN exist from both natural and anthropogenic sources.

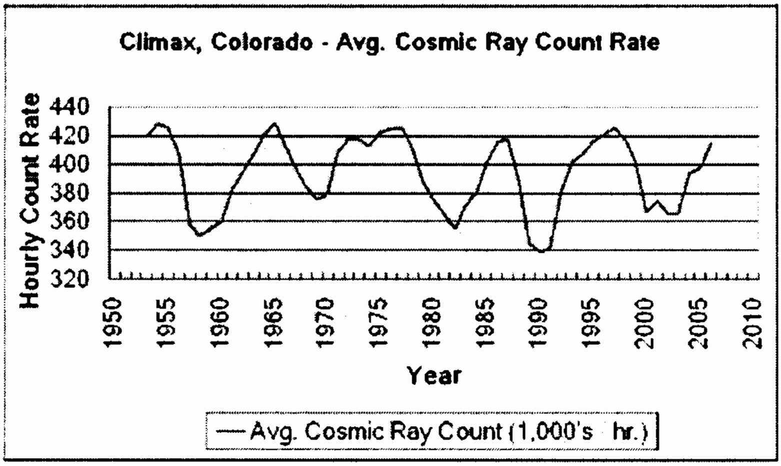


Fig. 6: Galactic cosmic ray intensity from 1954 to present day.

5. THE PALEOCLIMATE RECORD

Records of past environmental changes are preserved in paleo archives, such as ice caps, marine and lake sediments, trees, and long-lived corals. Reconstruction of these records requires that the properties measured in natural archives (proxies) be quantitatively translated into environmental parameters. In so doing, the proxies must be rigorously calibrated against direct observations, such as air temperature or ocean salinity. Thus, a period of overlap between the proxy record and contemporary data is important for calibrating paleo-reconstructions. Most natural archives contain many lines of evidence; for example, a single ice core contains indications of air temperature, atmospheric gas composition, volcanic activity, and dust deposition rates. Proxy records must be dated so that the timing of events, rates of change, and relationships between different archives can be established. Chronologies are based on a variety of methods. Several of these, such as radiocarbon dating, depend on radioactive decay. Others rely on counting of annual layers, whether of snow accumulation, tree rings, seasonally deposited sediments, or volcanic ash layers. Accurately dated coral terraces provide estimates for sea level history for about the last 150,000 years.

Over the past 500,000 years Earth's climate has varied cyclically between cold, glacial conditions and warm, interglacial periods. Warm, interglacial conditions have persisted for only short periods relative to the lengthy glacial ones. The cyclicity is driven by changes in the distribution of sunlight on the Earth's surface as the planet's orbit varies slightly through time. During glacial periods, the extent of ice on the land and surface ocean as well as variations in vegetation cover increased the reflectivity of the Earth's surface and reduced the amount of solar energy absorbed by the planet. Large parts of the Earth were drier, resulting in a dustier atmosphere, and more dust was deposited onto the ice sheets of Antarctica and Greenland and onto the surface of the ocean (Lambert et al., 2008). The glacial-age atmosphere was strikingly depleted of the greenhouse gases CO_2, CH_4, and N_2O.

Close correspondence among changes in ice volume, sea-surface temperature, dust deposition, and greenhouse gas concentrations occurred. Detailed comparisons of the relative timing of these variations yield considerable insight into climate feedbacks and the dynamics of the interconnections.

5.1 Glacial-interglacial cycles

Periodic changes in the Earth's orbit around the Sun and in the tilt (obliquity) of the Earth's axis control the seasonal and latitudinal distribution of incoming solar radiation (Hays et al., 1976; Berger, 1978). The time of year when the Earth is closest to the Sun varies with quasi-periodicities of about 19,000 and 23,000 years. The obliquity of the Earth's axis varies between 22° and 24.5° with a quasi-periodicity of about 41,000 years. When the tilt is greater, the poles are exposed to more sunlight.

Milutin Milankovic in the 1930s was the first to explain the cause of ice ages. He argued that glaciation occurs when solar insolation intensity is weak at high northern latitudes in summer. This occurs when both Earth's spin axis is less tilted with respect to the orbital plane and the aphelion (the point of Earth's orbit that is farthest from the Sun) coincides with summer in the

Northern Hemisphere. When there is less insolation during summer, snow and ice persist through the year, gradually accumulating into an ice sheet. The trigger for an ice age depends on the intensity of Northern Hemisphere summer sun (i.e., whether the solar intensity is large enough to melt the ice that accumulates over winter). If the solar flux < 475 W/m^2 at 65 °N, then this is insufficient to melt the ice.

The forcing induced by orbital variations alone is not sufficient to produce the ice ages; climate feedbacks are essential to produce the massive, synchronous glaciation in both hemispheres (Lorius et al., 1990). At the onset of an ice age, the small initial cooling resulting from the orbital changes is then amplified as the CO_2 concentration falls. At higher levels of CO_2, this trigger threshold decreases because the temperature is higher, i.e. an ice age is less likely. The onset of the last ice age, about 116,000 years ago, corresponded to a 65°N mid-June insolation about 40 W/m^2 lower than today. Earth would not naturally enter another Ice Age for 30,000 years. An outstanding question in describing the onset of ice ages is the mechanism by which the CO_2 level was driven down. There is general agreement that the CO_2 decrease is related to changes in the carbon uptake by the oceans, with some theories relying on enhancement of the biological pump, but the actual mechanism has yet to be firmly identified.

A predominant feature of the ice ages is their asymmetry, that is a long period of cooling is (~ 90,000 years) terminated by a dramatic period of warming (~ 10,000 years). In view of the weak orbital forcings, such a dramatic climate change must involve both substantial feedbacks, and, in the case of the ice ages, these involve both surface albedo and GHGs. Ice sheet disintegration can proceed rapidly, spurred by the large change in absorbed solar energy when ice and snow become wet. As solar insolation slowly increases due to orbital variations, the first melt in the spring is pushed earlier, leading to a longer period of wet snow/ice surface. Decreased surface albedo and ice sheet loss is accompanied by release of GHGs, leading to a relatively rapid termination of the ice age. The albedo feedback mechanism also explains the slow onset of glaciation.

The Vostok ice core in East Antarctica has provided a wealth of information on climate conditions; in 1998 the ice core reached a depth of 3623 m, corresponding to ~ 420,000 years back in time (Petit et al., 1999). The ice core allows an examination of each glacial commencement and termination over the past 420,000 years (Figure 7). The sequence of events during the termination of an ice age is of special interest. Temperature, CO_2, and CH_4 increase in phase during a termination. Uncertainty in the phasing arises from the sampling frequency and uncertainty in gas-age/ice-age differences. In an analysis of the Vostok core, Fischer et al. (1999) concluded that CO_2 concentration increases lagged Antarctic warmings by 600 ± 400 years. Petit et al. (1999) were a bit more cautious in inferring the sign of the phase relationship between CO_2 and temperature at the start of glacial terminations. During a deglaciation, the atmospheric concentration of CH_4 rises slowly, then jumps to a maximum value during the last half of the deglacial temperature rise. The CH_4 jump corresponds to a rapid Northern Hemisphere warming and an increase in the in the rate of Northern Hemisphere deglaciation that occurs within 1000 years of the CH_4 jump. Overall, the ice core data suggest that the sequence of forcings during each glacial termination were: orbital forcing that produced a small amount of warming, followed by two strong amplifiers, with greenhouse gases acting first, and then further enhancement via ice-albedo feedback (Monnin et al., 2001; Caillon et al., 2003). The end of the deglaciation is characterized by a clear CO_2 maximum.

In summary, during a deglaciation, slow changes in orbital parameters caused greater amounts of summer sunlight to fall in the Northern Hemisphere. This relatively small change caused ice to retreat in the north, which decreased the albedo. The loss of reflecting surface led to further warming in a feedback effect. About 600 years or so after that process started, CO_2 and CH_4 concentrations began to rise, which amplified the warming trend even more, and GHG forcing took over as the dominant factor in the ultimate change. Thus, GHGs did not initiate the warming periods but acted as the amplifier. The 600-year time lag is about the amount of time required for the deep ocean to mix, allowing CO_2 stored in the deep ocean to be brought up to the surface where it can be released to the atmosphere. CO_2 is also released from warming soils and CH_4 from melting permafrost.

Western Equatorial Pacific sea surface temperature was about 3°C colder during the last ice age than today. Since this area of the ocean was relatively unaffected by changes in higher latitude ice cover and in ocean circulation, the cooling can be explained only in terms of changes in atmospheric GHGs. A number of hypotheses exist to explain the low CO_2 concentrations during glacial times. The ocean is the most important of the relatively fast-

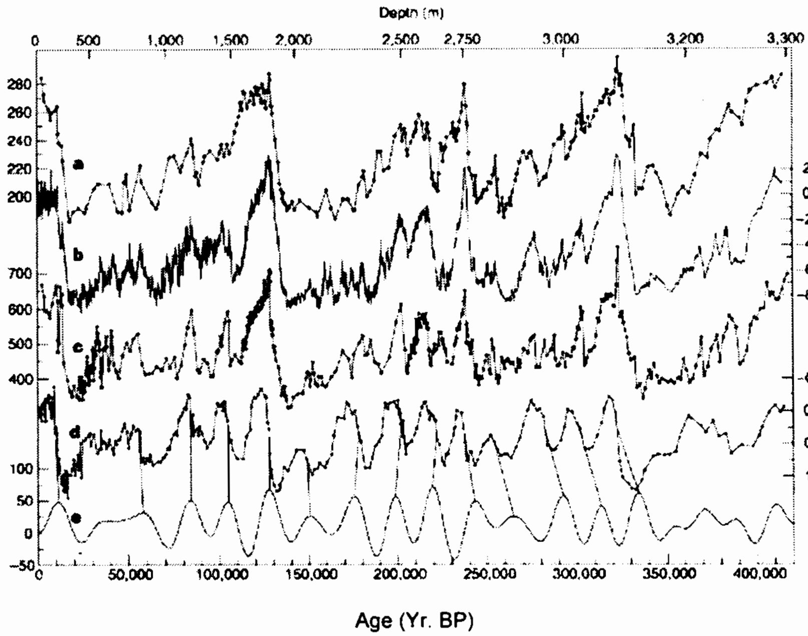


Fig. 7: Time series of data from the Vostok ice core (Petit et al., 1999). The lower axis gives time before present, and the upper axis shows the corresponding depth of the core. Curve **a** is CO_2 concentration (ppm); curve **b** is atmospheric temperature; curve **c** is CH_4 concentration (ppb); curve **d** is $\delta^{18}O$ of atmospheric O_2; curve **e** is mid-June insolation (W/m^2). $\delta^{18}O$ of O_2 reflects changes in ice volume and in the hydrological cycle.

exchanging (< 1000 year) carbon reservoirs. On these time-scales, atmospheric CO_2 is controlled by the interplay between ocean circulation, marine biological activity, ocean-sediment interactions, seawater carbonate chemistry, and air-sea exchange. CO_2 is more soluble in colder waters; so, changes in ocean temperature can alter atmospheric CO_2. The Southern Ocean is especially important in this regard because it is where large volumes of deep water masses are formed and where large amounts of biological nutrients upwell to the surface. Support for the role for the Southern Ocean in controlling atmospheric CO_2 during glacial times is provided by the concurrent decrease of Antarctic temperature and atmospheric CO_2.

Climate has varied over intervals much shorter than glacial cycles; paleoclimate records reveal that synchronous changes occurred on millennial timescales, especially during the ice ages. Strong evidence exists that these were the result of changes in the North Atlantic ocean circulation caused by fresh water flooding the ocean surface as ice sheets on the continents collapsed. During the last ice age and the transition to the present warm period (the Holocene) the Earth's climate exhibited swings that provide clues as to how atmosphere, ocean, and ice cover work together. Among the most prominent are the Dansgard/Oeschger (D/O) events, a series of rapid warmings, up to 16 °C in Greenland, recorded in ice cores. Even in regions of the Earth far from the North Atlantic, there is a clear atmospheric signature D/O events. Atmospheric CH_4 concentrations peaked during every warm episode. Methane concentrations are globally uniform and therefore can be used to synchronize high-resolution ice core records from Greenland and Antarctica. Inferred temperatures in Antarctica show a slow warming that preceded the abrupt warming in the Northern Hemisphere by 1000 - 2000 years. When the north finally warmed (within less than a century), a slow cooling began in the Southern Hemisphere. This out-of-phase warming and cooling between Greenland and Antarctica is a result of changes in the surface and deep water circulation of the Atlantic and associated exchanges of heat. In the modern ocean, the overall transport of heat is from the southern to the northern Atlantic. When the circulation in the north collapses, less heat is exported from the south, and the Southern Ocean warms. Re-establishment of the present-day circulation produces a D/O event and the south starts cooling again.

5.2 The Holocene

The transition from glacial to Holocene conditions, which occurred over about 12,000 years, was punctuated by several episodes of dramatic cooling (Lehmann and Keigwin, 1992; von Grafenstein et al., 1998). These periods were most likely the result of ocean circulation and sea surface temperature variations in the North Atlantic triggered by massive freshwater outflows from the Scandinavian and Laurentide ice shields. Contemporaneous changes in climate were recorded in sites ranging from Greenland, central Europe, and central North America.

The paleoclimate history teaches us that Earth's climate is remarkably sensitive to forcings through feedbacks that tend to be predominantly positive. The Holocene, now almost 12,000 years in duration, during which human civilization has developed, has been characterized by a remarkably stable climate. One argument that has been advanced to explain this stability is that the planet is warm enough for the great ice sheets of the Northern Hemisphere to be absent, but not warm enough for a disintegration of the Antarctica or Greenland ice sheets.

About 6,000 years ago, the Earth experienced a period of increased temperature, referred to as the Mid-Holocene Warming, during which temperatures were warmer than those today. The warm conditions occurred, however, only in the summer and only in the Northern Hemisphere. Orbital forcing was the cause of this event.

A period of generally warmer temperatures recorded mainly in western Europe between 1000 - 1200 A. D. has been referred to as the Medieval Warm Period (MWP). The concept of a Medieval Warm Epoch or High Medieval, centered around 1100 to 1200 A. D., was introduced by H. H. Lamb in 1965 based on historical anecdotes and paleoclimate data from western Europe (Bradley et al., 2003). Such conditions would have been associated with a prevailing anticyclonic circulation in summer and persistent westerly flow in winter. Since 1965, a number of paleotemperature series have been produced; nonetheless, well-calibrated data sets with decadal or higher resolution are still available only for a few dozen locations, only a few of which are from the tropics or the Southern Hemisphere. The limited database makes it difficult to determine if the warmer conditions in Europe were of global extent. Temperatures from 1000 to 1200 A.D. were almost the same (or 0.03 °C cooler) as those from 1901 to 1970. Although the extent of warming in the MWP is still uncertain, analysis of a variety of proxy records indicates that the geographic extent of warming in the late 20th century substantially exceeds that during the Medieval Warm Period (Osborn and Briffa, 2006). This period overlapped with a time of high solar activity, referred to as the Medieval Maximum. If solar irradiance was enhanced, changes in large-scale circulation patterns associated with the Arctic Oscillation may explain why Europe was relatively warmer in this period. Evidence for widespread hydrological anomalies (prolonged droughts in some regions and exceptional rains in others) are often associated with changes in atmospheric circulation regimes. The period from 1100 to 1260 A.D. was also characterized by increased volcanism. In the 20th century, such volcanic activity led to warm winters in northern Europe and northwestern Russian (Bradley et al., 2003). For an in-depth treatment of the climate of the MWP, the reader is referred to Hughes and Diaz (1994).

In terms of today's climate, whether or not the MWP was global in extent is irrelevant: the MWP was the result of increased solar activity, whereas the temperatures of the late 20th and early 21st centuries cannot be attributed to increased solar irradiance.

5.3 Paleocene-eocene thermal maximum

During the Paleocene-Eocene Thermal Maximum, 55 million years ago, tropical oceans warmed by 4 °C to 6 °C and high-latitude oceans by 8 °C to 10 °C in less than 10,000 years (Zachos et al., 2001); the event lasted for 50,000 to 200,000 years.

The cause of the PETM is still debated; the prevailing theory is that it was caused by release of massive amounts of CH_4 or CO_2 from thawing of methane clathrates in deep ocean sediments or a massive volcanic eruption that heated up coal deposits. The amount of temperature change in the PETM is consistent with a relatively high CO_2 climate sensitivity. According to our current understanding, such a large temperature change at the poles is associated with changes in ice cover; however, the absence of permanent sea or land ice at that time suggests the existence of another feedback that causes warming at high latitudes.

Since then, Earth has cooled; over the past million years the Earth has switched between ice ages and warmer interglacial periods. Only over the last 34 million years have CO_2 concentrations been low, temperatures relatively cool, and the poles glaciated. This long-term shift resulted, in part, from differences in volcanic emissions. Changes in chemical weathering of silicate rocks were also important. As CO_2 rises, temperature and precipitation increase and CO_2 weathering increases, producing a negative feedback. During the Paleocene-Eocene Thermal Maximum, global temperature increased by 5 °C in less than 10,000 years. At the same time it is estimated that 2000 Gt C as CO_2 entered the atmosphere and oceans.

Royer et al. (2007) have used the paleoclimate record of CO_2 levels over 420 million years to estimate the long-term equilibrium climate sensitivity. Their estimates are broadly consistent with those based on shorter-term climate records and indicate that a climate sensitivity greater than 1.5 °C for $2xCO_2$ has likely been a robust feature of the Earth's climate system for the last 420 million years. This conclusion is consistent with decadal to millennial records from the more recent past as well as millennial records from the ancient past, such as the Palaeocene-eocene Thermal Maximum.

5.4 Snowball earth

From 750 to 580 million years ago, the Earth was in an ice age more extreme than any since, termed Snowball Earth (Kirschvink, 2000; Hoffman and Schrag, 2002). Usually, the spread of ice produces further cooling by increasing the Earth's albedo. But ice on land eventually blocks the chemical weathering of rocks that removes CO_2 from the atmosphere; as a result, CO_2 accumulates, leading to warming. Snowball Earth may have been possible only because at that time in the Earth's history the continents were clustered on the Equator. In that case, CO_2 removal would have continued even as the ice sheets spread from the poles toward the equator. This trend would continue until most of the land was covered by ice, at which point CO_2 finally would start accumulating in the atmosphere.

6. TEMPERATURE CHANGES FROM PRE-INDUSTRIAL TO PRESENT

A great deal of analysis has been carried out on global temperature measurements. Figure 8 shows the analysis of the period 1880 - 2005 published by Hansen et al. (2006) based on land data, satellite measurements of sea surface temperature since 1982, and ship-based measurements in earlier years. Estimated 2σ (95% confidence) decreases from 0.1°C at the beginning of the 20th century to 0.05°C in recent decades. The current warming is nearly worldwide, generally larger over land than over ocean, and largest at high latitudes in the Northern Hemisphere. Overall warming was about 0.7°C between the late 19th century and year 2000. Slow warming, with large fluctuations, occurred over the century up to 1975, followed by rapid warming at a rate ~ 0.2°C per decade. Thus, the total warming from the late 19th century to year 2008 is about 0.8 °C. The largest warming has taken place over remote regions, especially high latitudes. Warming occurs over ocean areas, far from direct human effects. Warming over the oceans is less than that over land; this is the expected response to a forced climate change because of the large thermal inertia of the ocean. Three somewhat distinct periods are discernable in the temperature record in Figure 8:

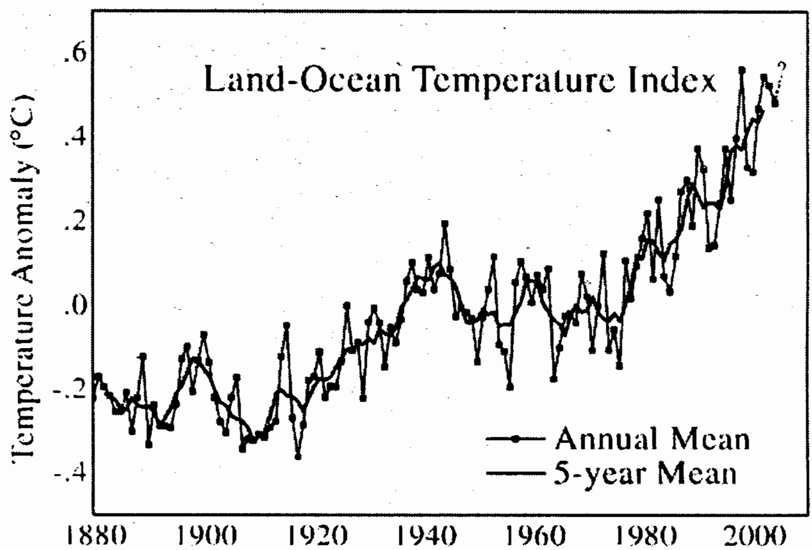


Fig. 8: Global temperature anomaly (°C) relative to 1951 - 1980 as derived from surface air measurements at meteorological stations and ship and satellite measurements (Hansen et al., 2006).

1910 – 1940	+0.35 °C
1940 – 1970	-0.1 °C
1970 – 2007	+0.55 °C

The three decades 1940 - 1970 were cooler than those preceding them. During this period sunspot activity and hence solar insolation was low. In addition, several volcanic eruptions occurred that led to enhanced stratospheric sulfate aerosol and its associated cooling. Also, emissions controls on fine particles were generally absent or weak prior to 1970; as a result, tropospheric aerosol levels were high, leading to additional cooling. Eleven of the 12 warmest years on record occurred since 1995. Based on satellite data from 1978 to 2002, the global troposphere (up to 10 km) has warmed at a rate of +0.22 °C to 0.26 °C per decade, consistent with the warming trend derived from surface meteorological stations (Vinnikov and Grody, 2003). The Antarctic winter troposphere temperature has increased at a rate of 0.5 °C to 0.7 °C per decade over the past 30 years (Turner et al., 2006). The stratosphere has cooled since 1979, owing to ozone depletion.

Greenhouse gases (CO_2, CH_4, N_2O, halocarbons) form a uniform blanket over the Earth, and their radiative forcing operates day and night. In addition, their atmospheric lifetimes are relatively long: CH_4 and N_2O have lifetimes of about 10 years and 120 years, respectively. CO_2 does not have an easily calculated lifetime, as it is removed by both the biosphere and the oceans over vastly different time scales (Joos et al., 1996, Shine et al, 2005). Of a unit pulse of CO_2:

50 % is removed in 30 years

30 % is removed in a few centuries

20 % is removed in 1000's of years (or remains permanently in the atmosphere due to permanent change in ocean pH).

Aerosols have a tropospheric lifetime measured in weeks; because of this, aerosol levels are highly non-uniform over the Earth, with highest concentrations occurring in regions of highest emissions (i.e. the Northern Hemisphere). Aerosols injected into the stratosphere, such as from a volcanic eruption, have a lifetime of about two years, owing to the absence of removal mechanisms (wet and dry deposition) present in the troposphere. Because aerosols affect solar radiation, their forcing occurs only during daylight hours.

Aerosols diminish the amount of solar radiation reaching the Earth's surface, so-called *global dimming*. While GHGs are more or less evenly distributed over the entire globe, aerosols are disproportionately concentrated in the Northern Hemisphere. Based on this, one might expect proportionately more warming in the Southern Hemisphere, but that is the opposite of what is observed. The explanation lies in the fact that uniform CO_2 concentrations do not imply uniform heating. Dynamical effects (changes in winds and ocean circulation) can exert as large an impact, locally, as GHG-induced forcing. Because the Northern Hemisphere contains disproportionately more land than the Southern Hemisphere, GHG-induced heating disproportionately affects the Northern Hemisphere. The ocean absorbs more heat without warming nearly as much, since it distributes heat rapidly in the upper layers via convection. Over land, most extra heat is transferred directly to the atmosphere. Another important factor is the difference in ocean dynamics between the Northern and Southern Hemisphere. Heat is mixed more efficiently into the deeper waters of the Southern Ocean. The interior of Antarctica has not warmed appreciably in the last few years. Thompson and Solomon (2002) showed that the Southern Annular Mode (a pattern of variability that affects the westerly winds around the Antarctic continent) has been in a more positive phase (stronger winds) in recent years; this acts as a barrier, preventing warmer mid-latitude air from reaching the continent, and may be a result of a combination of stratospheric O_3 depletion and stratospheric cooling owing to CO_2 (Shindell and Schmidt, 2004).

7. CLIMATE FORCING AND RESPONSE

The amount of excess energy the Earth is currently absorbing implies that the Earth is out of energy balance (solar energy in − infrared energy out > 0). As we have noted, a time lag exists before the Earth's climate can respond fully to the changes in forcing; this implies that additional heating is "in the pipeline." The temperature "in the pipeline" is realized if the atmospheric GHG composition were to be held constant into the future. Because of the long thermal response time of the ocean, the atmosphere exists in a quasi-equilibrium as the ocean changes slowly.

Because the heat capacity of the land surface is so small compared to that of the ocean, any significant imbalance in the energy budget ultimately leads to heating of the oceans. Ocean heat content estimates have become much more accurate over the last decade owing to both satellite altimeter and surface data. One result of this increased knowledge of the ocean heat content is that this quantity can now serve as a means to check model predictions.

Under equilibrium climate conditions the solar heat flux to the upper layer of the ocean is balanced by a heat flux from the ocean to the atmosphere, which establishes the temperature of the ocean surface layer. Heat transfer between the ocean and atmosphere is controlled by a thin layer of thickness O(1 mm) at the ocean surface through which heat is transferred by conduction. The temperature gradient across this thin conduction layer determines the heat flux, and at equilibrium the input of solar energy is balanced by the heat flux to the atmosphere. Imagine that the GHG concentration in the atmosphere is suddenly increased. The absorption of infrared radiation by GHGs leads to a downward flux of infrared radiation, which penetrates only a few micrometers into the conduction layer at the ocean-air interface. The effect of the downward infrared flux is to alter the temperature gradient across the conduction layer so that the surface temperature is slightly higher; the result is that the flux of heat through the layer to the atmosphere is reduced. Thus, more of the energy acquired by the bulk of the ocean surface layer from absorption of solar radiation remains in the ocean, leading to an increased temperature of the upper layer of the ocean. Short-term absorption of heat is concentrated in the upper 90 m or so of the ocean. Most of the increase in ocean temperature is confined to the top 300 to 700 m. The timescale for transport from the surface layer into the deeper ocean is decades to centuries, so this response is quite as expected.

7.1 Radiative forcing and climate sensitivity

Radiative forcing and climate sensitivity are the key concepts in understanding global mean temperature response. A simple globally averaged linear forcing-feedback model has proved to be remarkably robust. Even in sophisticated general circulation models, global mean radiative forcing is a useful predictor of global mean temperature response (National Research Council, 2005). When the climate system is subjected to a radiative forcing, ΔQ, a net flux imbalance, N, will exist until the system reaches a new equilibrium. For all practical purposes, the net flux imbalance is the excess energy stored in the oceans. If the globally average surface temperature change in response to the forcing is ΔT, the energy balance can be expressed as (Gregory et al., 2004),

$$N = \Delta Q - \lambda^{-1}\,\Delta T \tag{1}$$

where λ is the climate sensitivity (°C / W/m^2). Equation (1) includes only climate feedbacks that are proportional to temperature, i.e. water vapor clouds, surface albedo, atmospheric temperature profile (lapse rate).[6] Conceptually, the Earth's energy balance is perturbed by a forcing ΔQ, in turn leading to a climate response, ΔT. If ΔQ is held constant, the Earth eventually reaches a new equilibrium at which $N = 0$, that is, the extra heat content of the oceans is zero, as all the forcing is manifest in the new equilibrium temperature.

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[6] The temperature changes implicit in Equation (1) are those governed largely by the ocean mixed layer relaxation time. Thus, this equation may not reflect the impact of the more slowly responding elements of the climate system, such as ice sheets or the carbon cycle itself.

The benchmark forcing scenario that is generally considered is a doubling of atmospheric CO_2 (from 280 ppm to 560 ppm). At the new equilibrium corresponding to $2xCO_2$,

$$\Delta Q_{2xCO_2} = \lambda^{-1}\Delta T_{2xCO_2} \tag{2}$$

and the climate sensitivity is

$$\lambda = \left(\frac{\Delta T_{2xCO_2}}{\Delta Q_{2xCO_2}}\right) \tag{3}$$

Then Equation (1) can be written as

$$N = \Delta Q - \left(\frac{\Delta Q_{2xCO_2}}{\Delta T_{2xCO_2}}\right)\Delta T \tag{4}$$

The forcing from $2xCO_2$ is $\Delta Q_{2\,x\,CO_2} = 3.7$ W/m^2. At present day, the net heat imbalance in the oceans is estimated as (Hansen et al., 2005)

$N = 0.85 \pm 0.15$ W/m^2

and the global-mean temperature increase from pre-industrial to 2000 is $\Delta T =$ 0.7 °C. Then Equation (4) becomes

$$\Delta Q = \frac{2.6}{\Delta T_{2xCO_2}} + 0.85 \tag{5}$$

This simple energy balance model shows the inverse relationship between the temperature increase predicted at $2xCO_2$ and the total anthropogenic forcing. A range of predictions of $\Delta T_{2\,x\,CO_2}$ that generally encompasses those of climate models is 1.5 °C to 4.5 °C. At the lower end of this range, the forcing consistent with Equation (5) is $\Delta Q = 2.6$ W/m^2; at the upper end, ΔQ =1.43 W/m^2. The relationship in Equation (5) is shown in Figure 9.

The net direct climate forcing due to aerosols, as estimated by IPCC (2007) and broken down by source is:

Sulfate	-0.4 ± 0.2 W/m^2
Organic carbon	-0.05 ± 0.05
Black carbon[7]	$+0.2 \pm 0.15$
Biomass burning	-0.03 ± 0.12
Nitrate	-0.1 ± 0.1
Mineral dust	-0.1 ± 0.2
Total direct	-0.5 ± 0.4 W/m^2

The indirect climate forcing (effect of aerosols on cloud albedo) was estimated as -0.7 [-1.1, + 0.4]W/m^2, and no estimates were given of the forcing attributable to the effect of aerosols on cloud lifetime. Thus, the total forcing was estimated as -1.2 W/m^2. This can be compared with the total forcing due to GHGs and tropospheric O_3 of +3.0 W/m^2. Thus, according to the IPCC estimate, aerosols reduce GHG + O_3 forcing by 40%, globally.

Since the forcing from GHGs is well constrained, the implication of Equation (5) is that if present-day negative aerosol forcing is estimated to be larger, the climate sensitivity consistent with the observed temperature change must itself be larger. This relationship can be shown in somewhat different form than that in Figure 9, namely that in which ΔT_{2xCO_2} is a function of the assumed net aerosol forcing (Figure 10). If net aerosol forcing (negative) exceeds -1.2 W/m^2, then the amount of "cancellation" of GHG warming is even greater; consequently, the implied climate sensitivity to produce the actual observed temperature increase must be larger. At a

[7]Like all aerosols, black carbon scatters a portion of an incoming solar beam back to space, which leads to a reduction in solar radiation reaching the Earth's surface. A portion of the incoming solar beam is also absorbed by the black carbon. The black carbon absorbs radiation yet again from the diffuse upward beam of scattered sunlight, reducing the solar radiation reflected back to space. (This effect is particularly accentuated when absorbing aerosols lie above clouds.) With a sufficiently absorbing aerosol, the cooling caused by radiation scattering back to space can be compensated for by the heating caused by absorbed radiation, leading to no net change in the radiative balance at the top of the atmosphere. There is little disagreement that the rise of aerosols since pre-industrial times has led to both a substantial reduction in solar radiation at the surface and increased solar heating of the atmosphere itself. But global models disagree as to the magnitude of these effects.

present-day aerosol forcing of −2.0 W/m^2, for example, the climate sensitivity to 2xCO_2 must be almost 4.5 °C; at a net aerosol forcing of −1.0 W/m^2, the 2xCO_2 sensitivity is close to 1.5° C. The shaded area in Figure 10 represents the climate sensitivity range of 1.5 to 4.5 °C for 2xCO_2. A range of aerosol forcings from about −0.5 to −2.0 W/m^2 encompasses the range of climate sensitivities from 1.5 °C to 4.5 °C. At the IPCC (2007) estimate of total aerosol forcing of −1.2 W/m^2, the GISS Model E, which was used to produce Figure 10, predicts a climate sensitivity of 2° C for 2xCO_2. Kiehl (2007) provides additional discussion on the interplay between aerosol forcing and climate sensitivity.

Uncertainty in global forcing attributable to aerosols, therefore, translates into an uncertainty in climate sensitivity based on the observed change in global temperature over the last few decades. If the value of present-day aerosol (negative) forcing is deemed to be larger, the inferred climate sensitivity also gets larger; in other words, if the cooling effect of aerosols is larger, the sensitivity of climate change (the amount of warming as a result of a change in forcing) must be larger in order to be consistent with the observed temperature rise.

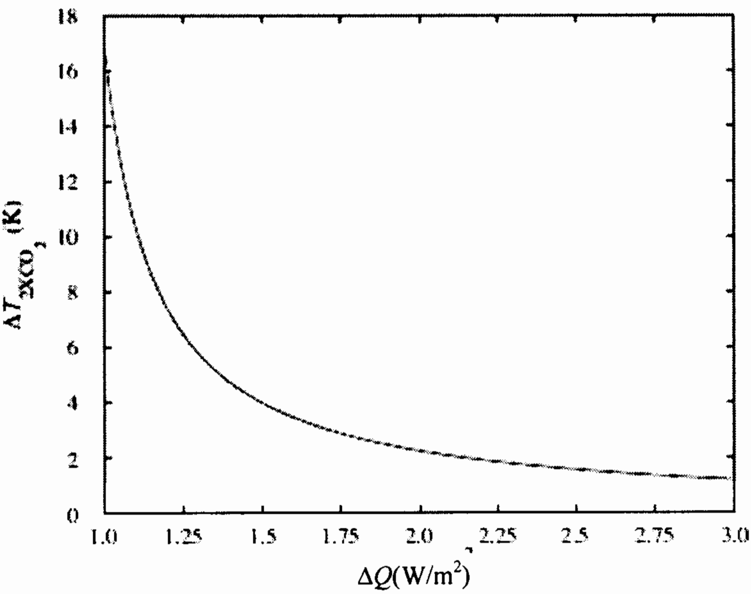

Fig. 9: Relation between temperature increase due to a doubling of CO_2 (from 280 ppm to 560 ppm) and net radiative forcing of the climate system, as described by Equation (5).

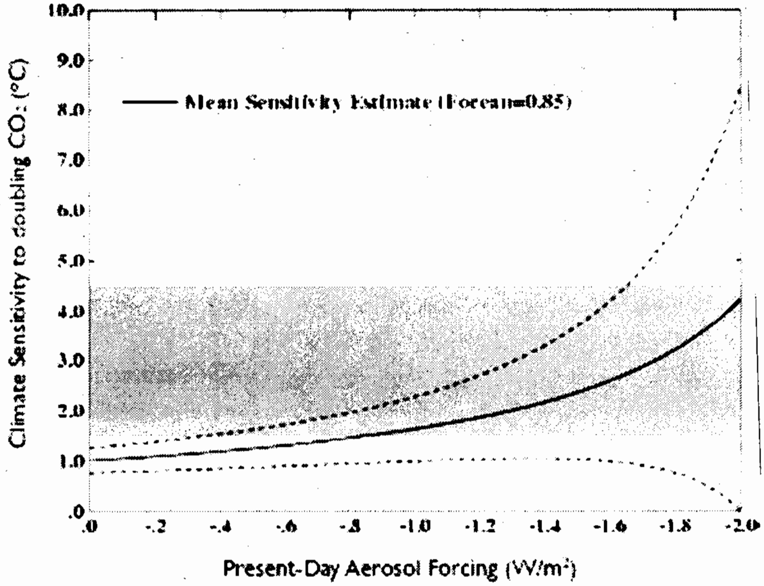

Fig. 10: Climate sensitivity (°C) as a function of assumed present-day net aerosol forcing (W/m^2), as computed by Goddard Institute for Space Studies (GISS) Model E for the 20th century (Gavin Schmidt, personal correspondence).

7.2 Attribution of climate change

From the study of past climate it is clear that many factors influence climate: solar activity, oscillations in Earth's orbit, greenhouse gases, ice cover, vegetation on land (or the lack of it), the configuration of the continents, dust thrown up by volcanoes or wind, the weathering of rocks, etc. Complex interactions between many of these factors amplify or dampen changes in temperature. The critical question we face is what is the cause of the current, rapid warming?

The culmination of the attribution of observed climate change is to simulate the climate of the last century or so using all known and quantifiable forcings. Some of the forcings are well established (e.g. well-mixed GHGs, volcanic emissions, solar irradiance), while others are less certain (tropospheric aerosol effects, land use changes). Given the uncertainties, best estimates are made consistent with observations of the actual forcing agents. A number of independent groups have carried out such simulations; the test is the extent to which the simulations reproduce the elements of climate change over the period. Figure 11 shows the results of Hansen et al. (2005). See also Karoly et al. (2003). The overall surface temperature is closely represented in this (and other studies, as well). In view of uncertainties in the forcings together with compensating errors in the climate sensitivity, evaluating both predicted temperature change and predicted ocean heat content allows one to establish the net radiation imbalance and to demonstrate that the models are consistent with both the surface and ocean changes.

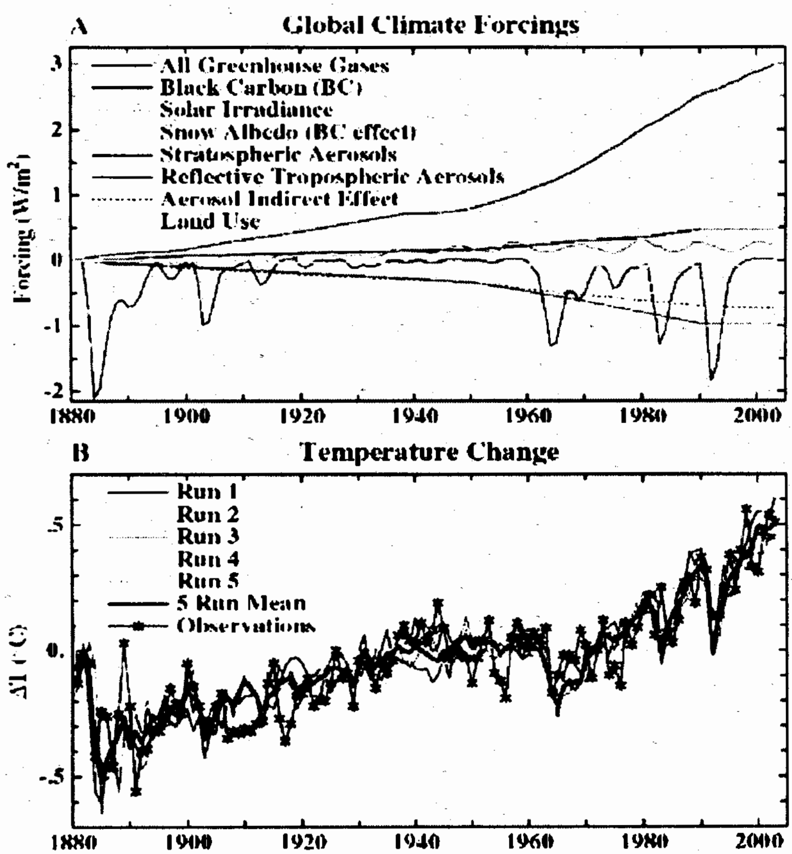

Fig. 11: Global climate forcings over the period 1880 - 2005 (Panel A) and corresponding predicted and observed temperature change (Panel B) (Hansen et al., 2005).

Despite the uncertainty inherent in comparisons such as that in Figure 11, since the surface temperatures and ocean heat content are rising together, this rules out intrinsic variability of the climate system as the cause of the recent warming. This is the case because internal climate changes, such as ENSO or thermohaline variability, are related to transfer of heat *within* the climate system, and therefore would occur only if energy is transferred from another reservoir (i.e. the ocean), which itself would need to be cooling. One must conclude that our understanding of forcings and long-term observations of land surface and ocean temperature changes is consistent within the range of uncertainties.

The current energy imbalance of 0.85 ± 0.15 W/m^2 implies a further warming of about 0.6 °C even if GHG concentrations were immediately stabilized at their present values.

7.3 General Circulation Models

General circulation models (GCMs) are based on the laws of physics, including couplings of the circulations of atmosphere and oceans, along with descriptions of the feedbacks between all components of the climate system, including the cryosphere and biosphere. General circulation models have been able to reproduce the main features of the current climate, the main features of the Holocene (the last 6,000 years) and the last Glacial Maximum (21,000 years ago).

Uncertainty in representation of physical processes at the spatial scale of general circulation models is the main source of differences in predicted climate sensitivity among models. In a sensitivity study designed to explore the range of variation of models, Murphy et al. (2004) varied 29 GCM parameters, one by one, and analyzed the results of 20-year simulations under present day and $2xCO_2$ conditions. They computed probability density functions of climate sensitivity; in the 5 - 95% probability range, warming ($2xCO_2$) is between 2.4 °C and 5.4 °C. Even with the relatively wide range of systematic parameter variation, the fundamental responses of the climate models lie within the range derived from paleoclimate data.

The range of variation of equilibrium temperature increases predicted for $2xCO_2$ is a result of the different ways in which climate models simulate processes internal to the climate system that either amplify or dampen the climate system's response to external forcing (See Appendix). Those climate feedbacks that directly affect the top-of-the-atmosphere radiation budget and do not involve chemical or biochemical processes in the biosphere or oceans involve water vapor, clouds, tropospheric temperature gradient (so-called temperature lapse rate), and surface albedo in snow/ice regions. The temperature lapse rate in the troposphere (i.e. the rate of decrease of temperature with altitude) affects the atmospheric transmission of infrared radiation to space; the steeper the decrease of temperature with height, the larger the greenhouse effect. Clouds strongly modulate the Earth's radiation budget, and the response of clouds to a global temperature change is a substantial part of the overall feedback effect. Of the global climate feedbacks, the water vapor feedback is the strongest; the magnitudes of these in current models can be summarized as follows (See Appendix): Water vapor feedback amplifies global mean temperature response by roughly a factor of 2, lapse rate feedback dampens it by about 20%, cloud feedback amplifies it by 10% - 50%, and surface albedo feedback amplifies it by about 10%.

The range of climate sensitivity estimates among models results primarily from the spread in predicted cloud feedback, but also with a substantial contribution of the water vapor, lapse rate, and surface albedo feedbacks. Cloud feedback has long been known to be the largest source of uncertainty in climate model predictions (even without considering the effects of aerosol changes on clouds). In the tropics, large-scale overturning circulations are associated with intense deep convective regions and widespread cloud-free regions of sinking motion. In the extratropics, the atmosphere is characterized by large-scale disturbances. In the tropics, nearly all of the upward motion occurs within deep cumulus clouds, with gentle subsidence between clouds. At mid-latitudes, the atmosphere is organized in synoptic weather systems. The tropics and extratropics are characterized by a number of cloud types, from low-level boundary layer clouds to deep convective clouds. Because of the different cloud top altitudes and optical properties, the different cloud types affect the radiation budget differently. A model must reflect how a change in climate may affect the distribution of the different cloud types and their radiative properties. The distribution of cloud types is controlled by the large-scale atmospheric circulation as well as the surface boundary conditions, wind shear, etc. In the tropics, atmospheric dynamics control to a large extent changes in cloudiness. Boundary layer cloud amount is strongly related to the cloud types present, which depend on a number of factors. In the midlatitude regions clouds are controlled by the large-scale atmospheric dynamics.

Global cloud feedback tends to be positive in all climate models, but large intermodal differences exist; the key question is – what is the reason for the spread of global cloud feedback estimates? The frequencies of occurrence of different cloud types can be quite different as can the water content of the clouds. Differences in cloud feedbacks in areas dominated by low cloud responses can make a large contribution to the variance of the global feedback.

With respect to the water vapor and lapse rate feedback, at low latitudes GCMs predict a larger warming at altitudes than near the surface, leading to a negative lapse rate feedback. At mid and high latitudes, by contrast, models predict a larger warming near the surface, i.e. a positive lapse rate feedback. On a global average, the tropical lapse rate response dominates, and the climate change lapse rate feedback is negative. Although all GCMs predict this response, the magnitude of this feedback varies among models. Intermodel differences in global lapse rate feedbacks are primarily a result of different meridional patterns of surface warming; the larger the ratio of tropical to global warming, the larger the negative lapse rate feedback.

Despite the range of climate sensitivities predicted by general circulation models, this uncertainty does not invalidate the use of GCMs to projecting future climate. Most emissions scenarios for the next 50 to 100 years involve a substantial increase in GHG forcing. And, given that most estimates of aerosol changes over the next century will not be proportionately as large as that of GHGs (and even for some substances, decreasing), future forcing will be dominated by GHGs. Therefore, net negative aerosol forcing is expected to assume an increasingly less important role as a mitigating factor in global warming.

8. ICE

As the poles warm, interior regions of Greenland and Antarctica can be expected to receive more snowfall, which adds to the interior ice sheets, whereas the ice in contact with the ocean is predicted to erode. Mass changes of the Greenland ice sheet can be deduced from satellite gravity observations. From 2003 to 2005, the ice sheet lost 101 ± 16 gigaton (Gt)/year, with a gain of 54 Gt/year above 2000 m and loss of 155 Gt/year at lower elevations (Luthcke et al., 2006). The overall rate of ice sheet loss reflects a considerable change in trend (−113 ± 17 Gt/year) from a near balance during the 1990s. An independent measurement of Greenland ice sheet loss by satellite radar interferometry indicated that the mass deficient in the last decade increased from 90 to 220 km^3/year (Rignot and Kanagaratnam, 2006).

Arctic sea ice extent typically attains a seasonal maximum in March and minimum in September. Over the course of the modern satellite record (1979 to present), sea ice extent has declined significantly in all months, with the decline being most pronounced in September (Stroeve et al., 2007). Monthly ice extent for September 2007 was 4.28 x 106 square kilometers, 23% smaller than the previous benchmark of 5.56 x 106 square kilometers set in September 2005. (This ice loss relative to September 2005 equates to an area the size of Texas and California combined.) Ice extent in September 2007 was 50% lower than conditions in the 1950s to the 1970s.

The ice pack contains a mixture of first-year ice and multiyear ice (ice that has survived for one or more melt seasons). In general, older ice is thicker than younger ice. Within the central Arctic Ocean, the coverage of old ice over this period declined by 88% and ice that is at least 9 years old essentially disappeared. This change toward younger ice translates to a decrease in mean thickness of ice over the Arctic Ocean from 2.6 meters in March 1987 to 2.0 meters in 2007.

Sea surface temperatures (SSTs) over the Chukchi and East Siberian seas have increased since the year 2000. During the summer of 2007, SSTs over parts of these seas reached more than 3.5 °C. After the ice thins to a more vulnerable state in response to rising sea surface temperatures, rapid decay of the remaining summer ice cover can then ensue due to the albedo feedback mechanism.

All models evaluated in the Intergovernmental Panel on Climate Change Fourth Assessment Report show declining September sea ice from 1953 to 2006. As a group, however, they underpredict the observed trend (Stroeve et al., 2007). The reasons for this under-representation remain to be fully resolved, but overly thick ice in several of the models provides a partial explanation. Given these conservative model results, along with the unexpectedly large sea ice loss of 2007, the long-term outlook for sea ice is troubling.

9. CONCLUSIONS

We return to the questions posed at the outset.

1. What is the perturbation to the Earth's energy balance that has occurred as a result of the build-up of GHGs?

   The atmospheric concentrations of the long-lived greenhouse gases (CO_2, CH_4, N_2O, halocarbons) have increased substantially since pre-industrial times. That the source of atmospheric CO_2 is anthropogenic is substantiated both by records of fossil fuel use and by measurement of carbon isotope ratios in atmospheric CO_2. The radiative perturbation to the Earth's energy balance that has resulted from the increase in GHG levels can be computed quite precisely. At present, the total forcing due to the long-lived GHGs is +2.63 W/m^2; this quantity can be compared to the annual, global average absorbed solar insolation of 235 W/m^2. If GHG concentrations were to be held constant at present day levels, the Earth would eventually come to a new thermal equilibrium at which 235 W/m^2 of infrared radiation is emitted back to space. Because of the increased infrared absorption in the atmosphere, the Earth's overall temperature would have to increase; at present day levels of GHGs and aerosols, that temperature would be about 1.4 °C above the pre-industrial temperature.

2. Water vapor is the dominant greenhouse gas in the Earth's atmosphere; what role does it play in climate change?

   Water vapor is an internal climate forcing agent in the climate system, and, moreover, it is the most important greenhouse gas in the atmosphere. The atmospheric water vapor concentration adjusts rapidly (on the order of 10 days) to achieve a quasi-steady state between its largely oceanic source and its precipitation sink. As temperature increases, the absolute concentration of water vapor in the atmosphere increases in accordance with the Clasius- Clapeyron relation; a 1 °C temperature increase produces a 7% increase in concentration. If the Earth's temperature increases from an external radiative forcing such as an increase in solar insolation or an increase in CO_2, then the amount of water vapor in the atmosphere increases. That increase is accompanied by a further increase in absorption of Earth's infrared radiation, which leads to additional warming, via a positive feedback. The purely radiative warming from a doubling of CO_2 from its pre-industrial level of 280 ppm to 560 ppm would be 1.2 °C; the actual warming that would result is estimated to lie in the range of 1.5 °C to 4.5 °C. For a mid-range estimate of 3 °C, the factor of 2.5 amplification of the equilibrium temperature is a result of the positive feedback operating on the purely radiative CO_2 signal largely by H_2O vapor.

3. To what extent are changes in solar irradiance and the galactic cosmic ray flux responsible for the Earth's temperature increase over the past century?
   Solar insolation has remained essentially constant over the past 5 to 6 decades and therefore cannot explain the degree of warming that has occurred over that time. For galactic cosmic rays to be influential on climate, a necessary, but not sufficient, condition is that a net downward trend in the cosmic ray flux would be necessary; no such trend has been observed over this period. Although the small changes in solar forcing due to the Sun's 11-year sunspot cycle do not exert a net long-term influence on climate, the transient response to the modest change in forcing associated with these cycles can be discerned in the Earth's temperature record. Because both the magnitude of the forcing and the temperature response of the Earth to the sunspot cycles are known with precision, a climate sensitivity can be deduced; that sensitivity is about 1 °C / W/m^2, a value near the higher end of current calculated climate sensitivities.
4. What does the paleoclimate record reveal about the response of the Earth's climate to forcings?
   The paleoclimate record is a valuable source of information on how Earth's climate has changed in the past and the causes of those changes. The paleoclimate history reveals that Earth's climate is remarkably sensitive to forcings through positive feedbacks that act to amplify the climate change (either warming or cooling). The glacial-interglacial cycles, especially, show the profound importance of feedbacks associated with both GHG levels and the Earth's albedo. Climate sensitivities deduced from paleoclimate records are comparable to those calculated for more recent forcings and responses.
5. Aerosols (airborne particles) exert a cooling effect on climate; what role do they play in the observed temperature change of the planet?
   On the whole, aerosols exert a cooling effect on climate. The magnitude of the total radiative forcing due to aerosols represents the most uncertain component of net global forcing of all substances, and, of the total aerosol

forcing, that attributable to the effect of aerosol changes on cloud properties and lifetime is the most uncertain. The warming of climate is a result of a global positive forcing, which can be viewed as the net forcing from GHGs and aerosols. GHG forcing is calculated with great precision; thus, the uncertainty in net forcing to which the observed temperature increase is responding translates into an uncertainty in Earth's climate sensitivity. For a larger aerosol forcing, the Earth's climate sensitivity must be correspondingly greater. A larger climate sensitivity implies that, as future GHG levels continue to increase and positive GHG forcing gets larger and larger relative to the magnitude of negative aerosol forcing, future temperature rise will be larger. A significant research need is to determine more accurately global aerosol forcing.

6. And, finally, to what causes can the warming experienced by the Earth over the last century, and the last three decades in particular, be attributed?

   The global temperature increase over the last century, and the last three decades, in particular, is well outside of that which could be attributed to natural climate fluctuations; only a substantial positive climate forcing (a perturbation to the Earth's energy balance that results in more absorption of solar incoming radiation than emission of infrared outgoing radiation) could produce such an increase. From a careful analysis of all evidence, the inescapable conclusion is that no natural climate factor (e.g. solar insolation changes, volcanic outgassing) can be the cause, only the substantial increase in atmospheric CO_2 and other greenhouse gases.

APPENDIX – CLIMATE FEEDBACKS

At equilibrium, the net absorbed incoming solar insolation equals the net outgoing terrestrial longwave radition, both expressed in units of W/m^2. If an external perturbation, such as a change in the solar constant or a change in the atmospheric concentration of CO_2, is imposed on the climate system, the Earth's radiation balance is disequilibrated by an amount ΔQ, called a radiative forcing. The climate system responds to this radiative imbalance by changing the global mean temperature. At any time, the change in global mean temperature from its unperturbed equilibrium value, ΔT, can be related to the imposed radiative forcing and to the radiative imbalance N by

$$N = \Delta Q + \beta\, \Delta T \tag{A.1}$$

where β is a feedback parameter. We are expressing the energy balance somewhat differently than in Equation (1); namely we define β as a feedback parameter (W/m^2/°C). Also, for a positive ΔQ (i.e. warming), as expressed in Equation (A.1), $\beta < 0$. The reason we do this is that it facilitates deriving an expression that allows separation of the effects of individual feedback processes. The climate system reaches a new equilibrium when $N = 0$, and ΔT at $N = 0$ is the temperature change from the original equilibrium temperature to the new equilibrium.

Let x be a vector representing a set of climate variables that affect the Earth's outgoing infrared radiation. The feedback parameter β can be formally defined by

$$\beta = \frac{\partial N}{\partial T} = \sum_i \frac{\partial N}{\partial x_i}\frac{\partial x_i}{\partial T} + \ldots \tag{A.2}$$

The most fundamental feedback is the temperature dependence of the infrared emission rate according to the Stefan-Boltzmann law of blackbody emission. The actual global mean temperature response of the climate system can be compared to that which would occur if temperature were the only variable to respond to the radiative forcing, ΔT_{SB}, where β_{SB} is the feedback parameter that accounts only for the Stefan-Boltzmann response. The global mean temperature change that actually occurs when all the climate variables x_i respond to the change in T can then be expressed as

$$\Delta T = \left(\frac{\beta_{SB}}{\beta}\right) \Delta T_{SB} \tag{A.3}$$

The feedback parameter β can be considered to be (to first-order) the sum of the Stefan-Boltzmann (SB) response and those of all other feedbacks; then, one may write (neglecting interactions between feedbacks),

$$\beta = \beta_{SB} + \sum_{i \neq SB} \beta_i \tag{A.4}$$

where the summation is over all feedbacks other than SB.

Then the global mean temperature change can be expressed from (A.3) and (A.4) as

$$\Delta T = \left(\frac{\beta_{SB}}{\beta_{SB} + \sum_{i \neq SB} \beta_i} \right) \Delta T_{SB}$$

$$= \left(\frac{1}{1 + \sum_{i \neq SB} \frac{\beta_i}{\beta_{SB}}} \right) \Delta T_{SB} \qquad \text{(A.5)}$$

We define $g_i = -\beta_i / \beta_{SB}$ as the feedback gain for feedback i, and

$$g = \sum_{i \neq SB} g_i \qquad \text{(A.6)}$$

g is the sum of all feedback gains other than SB. Thus, Equation (A.5) can be written as

$$\Delta T = \left(\frac{1}{1 - g} \right) \Delta T_{SB} \qquad \text{(A.7)}$$

or simply,

$$\Delta T = f \Delta T_{SB} \qquad \text{(A.8)}$$

where f is the feedback factor, that is the factor by which the actual temperature change differs from the pure Stefan-Boltzmann response. If $g > 0$, $\Delta T > \Delta T_{SB}$.

For the Stefan-Boltzmann response, an increase in temperature leads to an increase in the longwave emission and thus reduces N. From the blackbody emission flux of σT^4, $\beta_{SB} = -4\sigma T^3$. If one assumes an emission temperature of the Earth of 255 K, then $\beta_{SB} = -3.8$ W/m^2/K. From GCM calculations, the more accurate value of β_{SB} is about -3.2 W/m^2/K (Soden and Held, 2006).

Individual climate feedbacks determined from GCM simulations have been estimated as follows (Bony et al., 2006):

	β_i (W/m^2/K)	f
Water vapor	1.80 ± 0.18	2.27
Lapse rate	-0.84 ± 0.26	0.8
Cloud	0.69 ± 0.38	1.28
Surface albedo	0.26 ± 0.08	1.09

Therefore, water vapor feedback amplifies the global temperature response by a factor of 2 or more. Lapse rate feedback acts to dampen a warming response by about 20%. The mean value of predicted cloud feedback acts to amplify warming by about 30%, and surface albedo feedback leads to about 10% amplification of warming.

If we consider these feedbacks acting in concert, Equation (A.7) predicts that

$$\Delta T = 2.5\ \Delta T_{SB}$$

and for 2xCO_2, ΔT_{SB} = 1.2 °C and ΔT = 3 °C, in essential agreement with a climate sensitivity deduced from the glacial-interglacial cycles.

GLOSSARY (IPCC, 2007)

Aerosols A collection of airborne solid or liquid particles, with a typical size between 0.01 and 10 μm that reside in the atmosphere for at least several hours. Aerosols may be of either natural or anthropogenic origin. Aerosols may influence climate in several ways: directly through scattering and absorbing radiation, and indirectly by acting as cloud condensation nuclei or modifying the optical properties and lifetime of clouds.

Albedo The fraction of solar radiation reflected by a surface or object, often expressed as a percentage. Snow-covered surfaces have a high albedo, the surface albedo of soils ranges from high to low, and vegetation-covered surfaces and oceans have a low albedo. The Earth's planetary albedo varies mainly through varying cloudiness, snow, ice, leaf area and land cover changes.

Albedo feedback A climate feedback involving changes in the Earth's albedo. It usually refers to changes in the cryosphere, which has an albedo much larger (~0.8) than the average planetary albedo (~0.3). In a warming climate, it is anticipated that the cryosphere would shrink, the Earth's overall albedo would decrease and more solar radiation would be absorbed to warm the Earth still further.

Anthropogenic Resulting from or produced by human beings.

Biosphere (terrestrial and marine) The part of the Earth system comprising all ecosystems and living organisms, in the atmosphere, on land (terrestrial biosphere) or in the oceans (marine biosphere), including derived dead organic matter, such as litter, soil organic matter and oceanic detritus.

Black carbon (BC) Operationally defined aerosol species based on measurement of light absorption and chemical reactivity and/or thermal stability; consists of soot, charcoal and/or possible light-absorbing refractory organic matter.

^{13}C Stable isotope of carbon having an atomic weight of approximately 13. Measurements of the ratio of $^{13}C/^{12}C$ in carbon dioxide molecules are used to infer the importance of different carbon cycle and climate processes and the size of the terrestrial carbon reservoir.

^{14}C Unstable isotope of carbon having an atomic weight of approximately 14, and a half-life of about 5,700 years. It is often used for dating purposes going back some 40 kyr. Its variation in time is affected by the magnetic fields of the Sun and Earth, which influence its production from cosmic rays.

Carbon cycle The term used to describe the flow of carbon (in various forms, e.g., as carbon dioxide) through the atmosphere, ocean, terrestrial biosphere and lithosphere.

Climate Climate in a narrow sense is usually defined as the average weather, or more rigorously, as the statistical description in terms of the mean and variability of relevant quantities over a period of time ranging from months to thousands or millions of years. The classical period for averaging these variables is 30 years, as defined by the World Meteorological Organization. The relevant quantities are most often surface variables such as temperature, precipitation and wind. Climate in a wider sense is the state, including a statistical description, of the climate system.

Climate feedback An interaction mechanism between processes in the climate system is called a climate feedback when the result of an initial process triggers changes in a second process that in turn influences the initial one. A positive feedback intensifies the original process, and a negative feedback reduces it.

Climate Feedback Parameter A way to quantify the radiative response of the climate system to a global surface temperature change induced by a radiative forcing (units: $W/m^2/°C$). It varies as the inverse of the effective climate sensitivity.

Climate system The climate system is the highly complex system consisting of five major components: the atmosphere, the hydrosphere, the cryosphere, the land surface and the biosphere, and the interactions between them. The climate system evolves in time under the influence of its own internal dynamics and because of external forcings such as volcanic eruptions, solar variations and anthropogenic forcings such as the changing composition of the atmosphere and land use change.

Cloud condensation nuclei (CCN) Airborne particles that serve as an initial site for the condensation of liquid water, which can lead to the formation of cloud droplets.

Cloud feedback A climate feedback involving changes in any of the properties of clouds as a response to other atmospheric changes. Understanding cloud feedbacks and determining their magnitude and sign require an understanding of how a change in climate may affect the spectrum of cloud types, the cloud fraction and height, and the radiative properties of clouds, and an estimate of the impact of these changes on the Earth's radiation budget.

Cloud radiative forcing Cloud radiative forcing is the difference between the all-sky Earth's radiation budget and the clear-sky Earth's radiation budget (units: W/m^2).

Cryosphere The component of the climate system consisting of all snow, ice and frozen ground (including permafrost) on and beneath the surface of the Earth and ocean.

Dansgaard-Oeschger events Abrupt warming events followed by gradual cooling. The abrupt warming and gradual cooling is primarily seen in Greenland ice cores and in palaeoclimate records from the nearby North

Atlantic, while a more general warming followed by a gradual cooling has been observed in other areas as well, at intervals of 1.5 to 7 kyr during glacial times.

Detection and attribution Climate varies continually on all time scales. Detection of climate change is the process of demonstrating that climate has changed in some defined statistical sense, without providing a reason for that change. Attribution of causes of climate change is the process of establishing the most likely causes for the detected change with some defined level of confidence.

El Niño-Southern Oscillation (ENSO) The term El Niño was initially used to describe a warm-water current that periodically flows along the coast of Ecuador and Perú, disrupting the local fishery. It has since become identified with a basin-wide warming of the tropical Pacific Ocean east of the dateline. This oceanic event is associated with a fluctuation of a global-scale tropical and subtropical surface pressure pattern called the Southern Oscillation. This coupled atmosphere-ocean phenomenon, with preferred time scales of two to about seven years, is collectively known as the El Niño-Southern Oscillation (ENSO). It is often measured by the surface pressure anomaly difference between Darwin and Tahiti and the sea surface temperatures in the central and eastern equatorial Pacific. During an ENSO event, the prevailing trade winds weaken, reducing upwelling and altering ocean currents such that the sea surface temperatures warm, further weakening the trade winds. This event has a great impact on the wind, sea surface temperature and precipitation patterns in the tropical Pacific. It has climatic effects throughout the Pacific region and in many other parts of the world, through global teleconnections. The cold phase of ENSO is called La Niña.

Energy balance The difference between the total incoming and total outgoing energy. If this balance is positive, warming occurs; if it is negative, cooling occurs. Averaged over the globe and over long time periods, this balance must be zero. Because the climate system derives virtually all its energy from the Sun, zero balance implies that, globally, the amount of incoming solar radiation on average must be equal to the sum of the outgoing reflected solar radiation and the outgoing thermal infrared radiation emitted by the climate system. A perturbation of this global radiation balance, be it anthropogenic or natural, is called radiative forcing.

External forcing External forcing refers to a forcing agent outside the climate system causing a change in the climate system. Volcanic eruptions, solar variations and anthropogenic changes in the composition of the atmosphere and land use change are external forcings.

Global dimming Global dimming refers to perceived widespread reduction of solar radiation received at the surface of the Earth from about the year 1961 to around 1990.

Greenhouse effect Greenhouse gases effectively absorb thermal infrared radiation, emitted by the Earth's surface, by the atmosphere itself due to the same gases, and by clouds. Atmospheric radiation is emitted to all sides, including downward to the Earth's surface. Thus, greenhouse gases trap heat within the surface-troposphere system. This is called the greenhouse effect. Thermal infrared radiation in the troposphere is strongly coupled to the temperature of the atmosphere at the altitude at which it is emitted. In the troposphere, the temperature generally decreases with height. Effectively, infrared radiation emitted to space originates from an altitude with a temperature of, on average, –19 °C, in balance with the net incoming solar radiation, whereas the Earth's surface is kept at a much higher temperature of, on average, +14 °C. An increase in the concentration of greenhouse gases leads to an increased infrared opacity of the atmosphere, and therefore to an effective radiation into space from a higher altitude at a lower temperature. This causes a radiative forcing that leads to an enhancement of the greenhouse effect, the so-called enhanced greenhouse effect.

Greenhouse gas (GHG) Greenhouse gases are those gaseous constituents of the atmosphere, both natural and anthropogenic, that absorb and emit radiation at specific wavelengths within the spectrum of thermal infrared radiation emitted by the Earth's surface, the atmosphere itself, and by clouds. This property causes the greenhouse effect. Water vapor (H_2O), carbon dioxide (CO_2), nitrous oxide (N_2O), methane (CH_4) and ozone (O_3) are the primary greenhouse gases in the Earth's atmosphere. Moreover, there are a number of entirely human-made greenhouse gases in the atmosphere, such as the halocarbons and other chlorine- and bromine-containing substances, dealt with under the Montreal Protocol.

Halocarbons A collective term for the group of partially halogenated organic species, including the chlorofluorocarbons (CFCs), hydrochlorofluorocarbons (HCFCs), hydrofluorocarbons (HFCs), halons, methyl chloride, methyl bromide, etc. Many of the halocarbons have large Global Warming Potentials. The chlorine- and bromine-containing halocarbons are also involved in the depletion of the ozone layer.

Holocene The Holocene geological epoch is the latter of two Quaternary epochs, extending from about 11.6 ka to and including the present.

Ice age An ice age or glacial period is characterized by a long-term reduction in the temperature of the Earth's climate, resulting in growth of continental ice sheets and mountains glaciers (glaciation).

Ice core A cylinder of ice drilled out of a glacier or ice sheet.

Ice sheet A mass of land ice that is sufficiently deep to cover most of the underlying bedrock topography, so that its shape is mainly determined by its dynamics (the flow of the ice as it deforms internally and/or slides at its base). An ice sheet flows outward from a high central ice plateau with a small average surface slope. The margins usually slope more steeply, and most ice is discharged through fast-flowing ice streams or outlet glaciers, in some cases into the sea or into ice shelves floating on the sea. There are only three large ice sheets in the modern world, one on Greenland and two on Antarctica, the East and West Antarctic Ice Sheets, divided by the Transantarctic Mountains. During glacial periods there were others.

Insolation The amount of solar radiation reaching the Earth by latitude and by season. Usually insolation refers to the radiation arriving at the top of the atmosphere. Sometimes it is specified as referring to the radiation arriving at the Earth's surface.

Interglacials The warm periods between ice age glaciations. The previous interglacial, dated approximately from 129 to 116 ka, is referred to as the Last Interglacial.

Lapse rate The rate of change of an atmospheric variable, usually temperature, with height. The lapse rate is considered positive when the variable decreases with height.

Last Glacial Maximum (LGM) The Last Glacial Maximum refers to the time of maximum extent of the ice sheets during the last glaciation, approximately 21 ka. This period has been widely studied because the radiative forcings and boundary conditions are relatively well known and because the global cooling during that period is comparable with the projected warming over the 21st century.

Latent heat flux The flux of heat from the Earth's surface to the atmosphere that is associated with evaporation or condensation of water vapor at the surface; a component of the surface energy budget.

Little Ice Age (LIA) An interval between approximately AD 1400 and 1900 when temperatures in the Northern Hemisphere were generally colder than today's, especially in Europe.

Medieval Warm Period (MWP) An interval between AD 1000 and 1300 in which some Northern Hemisphere regions were warmer than during the Little Ice Age that followed.

Meridional Overturning Circulation (MOC) Meridional (north-south) overturning circulation in the ocean quantified by zonal (east-west) sums of mass transports in depth or density layers. In the North Atlantic, away from the subpolar regions, the MOC (which is in principle an observable quantity) is often identified with the Thermohaline Circulation (THC), which is a conceptual interpretation. However, it must be borne in mind that the MOC can also include shallower, wind-driven overturning cells such as occur in the upper ocean in the tropics and subtropics, in which warm (light) waters moving poleward are transformed to slightly denser waters and subducted equatorward at deeper levels.

Modes of climate variability Natural variability of the climate system, in particular on seasonal and longer time scales, predominantly occurs with preferred spatial patterns and time scales, through the dynamical characteristics of the atmospheric circulation and through interactions with the land and ocean surfaces. Such patterns are often called regimes, modes or teleconnections. Examples are the North Atlantic Oscillation (NAO), the Pacific-North American pattern (PNA), the El Niño-Southern Oscillation (ENSO), the Northern Annular Mode (NAM; previously called Arctic Oscillation, AO) and the Southern Annular Mode (SAM; previously called the Antarctic Oscillation, AAO).

Palaeoclimate Climate during periods prior to the development of measuring instruments, including historic and geologic time, for which only proxy climate records are available.

Pleistocene The earlier of two Quaternary epochs, extending from the end of the Pliocene, about 1.8 Ma, until the beginning of the Holocene about 11.6 ka.

Proxy A proxy climate indicator is a local record that is interpreted, using physical and biophysical principles, to represent some combination of climate-related variations back in time. Climate-related data derived in this way are referred to as proxy data. Examples of proxies include pollen analysis, tree ring records, characteristics of corals and various data derived from ice cores.

Quaternary The period of geological time following the Tertiary (65 Ma to 1.8 Ma). Following the current definition (which is under revision at present) the Quaternary extends from 1.8 Ma until the present. It is formed of two epochs, the Pleistocene and the Holocene.

Radiative forcing Radiative forcing is the change in the net, downward minus upward, irradiance (expressed in W/m^2) at the tropopause due to a change in an external driver of climate change, such as, for example, a change in the concentration of carbon dioxide or the output of the Sun. Radiative forcing is computed with all tropospheric properties held fixed at their unperturbed values, and after allowing for stratospheric temperatures, if perturbed, to readjust to radiative-dynamical equilibrium. Radiative forcing is called instantaneous if no change in stratospheric temperature is accounted for.

Reservoir A component of the climate system, other than the atmosphere, which has the capacity to store, accumulate or release a substance of concern, for example, carbon, a greenhouse gas or a precursor. Oceans, soils, and forests are examples of reservoirs of carbon. Pool is an equivalent term (note that the definition of pool often includes the atmosphere). The absolute quantity of the substance of concern held within a reservoir at a specified time is called the stock.

Response time The response time or adjustment time is the time needed for the climate system or its components to re-equilibrate to a new state, following a forcing resulting from external and internal processes or feedbacks. It is very different for various components of the climate system. The response time of the troposphere is relatively short, from days to weeks, whereas the stratosphere reaches equilibrium on a time scale of typically a few months. Due to their large heat capacity, the oceans have a much longer response time: typically decades, but up to centuries or millennia. The response time of the strongly coupled surface-troposphere system is, therefore, slow compared to that of the stratosphere, and mainly determined by the oceans. The biosphere may respond quickly (e.g., to droughts), but also very slowly to imposed changes.

Sea ice Any form of ice found at sea that has originated from the freezing of seawater. Sea ice may be discontinuous pieces (ice floes) moved on the ocean surface by wind and currents (pack ice), or a motionless sheet attached to the coast (land-fast ice). Sea ice less than one year old is called first-year ice. Multi-year ice is sea ice that has survived at least one summer melt season.

Sea level change Sea level can change, both globally and locally, due to (i) changes in the shape of the ocean basins, (ii) changes in the total mass of water and (iii) changes in water density. Sea level changes induced by changes in water density are called steric. Density changes induced by temperature changes only are called thermosteric, while density changes induced by salinity changes are called halosteric.

Sea surface temperature (SST) The sea surface temperature is the temperature of the subsurface bulk temperature in the top few meters of the ocean, measured by ships, buoys and drifters. From ships, measurements of water samples in buckets were mostly switched in the 1940s to samples from engine intake water. Satellite measurements of skin temperature (uppermost layer; a fraction of a millimeter thick) in the infrared or the top centimeter or so in the microwave are also used, but must be adjusted to be compatible with the bulk temperature.

Sensible heat flux The flux of heat from the Earth's surface to the atmosphere that is not associated with phase changes of water; a component of the surface energy budget.

Solar radiation Electromagnetic radiation emitted by the Sun. It is also referred to as shortwave radiation. Solar radiation has a distinctive range of wavelengths (spectrum) determined by the temperature of the Sun, peaking in visible wavelengths.

Stratosphere The highly stratified region of the atmosphere above the troposphere extending from about 10 km (ranging from 9 km at high latitudes to 16 km in the tropics on average) to about 50 km altitude.

Sunspots Small dark areas on the Sun. The number of sunspots is higher during periods of high solar activity, and varies in particular with the solar cycle.

Thermal expansion In connection with sea level, this refers to the increase in volume (and decrease in density) that results from warming water. A warming of the ocean leads to an expansion of the ocean volume and hence an increase in sea level.

Thermal infrared radiation Radiation emitted by the Earth's surface, the atmosphere and the clouds. It is also known as terrestrial or longwave radiation, and is to be distinguished from the near-infrared radiation that is part of the solar spectrum. Infrared radiation, in general, has a distinctive range of wavelengths (spectrum) longer than the wavelength of the red color in the visible part of the spectrum. The spectrum of thermal infrared radiation is practically distinct from that of shortwave or solar radiation because of the difference in temperature between the Sun and the Earth-atmosphere system.

Thermohaline circulation (THC) Large-scale circulation in the ocean that transforms low density upper ocean waters to higher-density intermediate and deep waters and returns those waters back to the upper ocean. The circulation is asymmetric, with conversion to dense waters in restricted regions at high latitudes and the return to the surface involving slow upwelling and diffusive processes over much larger geographic regions. The THC is driven by high densities at or near the surface, caused by cold temperatures and/or high salinities, but despite its suggestive though common name, is also driven by mechanical forces such as wind and tides. Frequently the name THC has been used synonymously with Meridional Overturning Circulation.

Total solar irradiance (TSI) The amount of solar radiance received outside the Earth's atmosphere on a surface normal to the incident radiation, and at the Earth's mean distance from the Sun. Reliable measurements of solar radiation can only be made from space and the precise record extends back only to 1978. The generally accepted value is 1,368 W/m^2 with an accuracy of about 0.2%. Variations of a few tenths of a percent are common, usually associated with the passage of sunspots across the solar disk. The solar cycle variation of TSI is of the order of 0.1%.

Tree rings Concentric rings of secondary wood evident in a cross-section of the stem of a woody plant. The difference between the dense, small-celled late wood of one season and the wide-celled early wood of the following spring enables the age of a tree to be estimated, and the ring widths or density can be related to climate parameters such as temperature and precipitation.

Troposphere The lowest layer of the atmosphere, from the surface to about 10 km in altitude at mid-latitudes (ranging from 9 km at high latitudes to 16 km in the tropics on average), where clouds and weather phenomena occur. In the troposphere, temperatures generally decrease with height.

Younger Dryas A period of 12.9 to 11.6 kya, during the deglaciation, characterized by a temporary return to colder conditions in many locations, especially around the North Atlantic.

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