Determination of a lower bound on Earth's climate sensitivity

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ABSTRACT.

Transient and equilibrium sensitivity or Earth's climate have been calculated using global temperature, forcing, and heating rate data for the period 1970-2010. We have assumed increased long wave radiative forcing in the period due to the increase of the long-lived greenhouse gases. By assuming that the change in aerosol forcing in the period to be zero we calculate what we consider to be lower bounds to these sensitivities, as the magnitude of the negative aerosol forcing is unlikely to have diminished in this period. The radiation imbalance necessary to calculate equilibrium sensitivity is estimated from the rate of ocean heat accumulation as 0.37 ± 0.03 W m^{-2} (all uncertainty estimates are 1-σ). With these data we obtain best estimates for transient climate sensitivity 0.39 ± 0.07 K (W $m^{-2})^{-1}$ and equilibrium climate sensitivity 0.54 $\pm$ 0.14 K (W $m^{-2})^{-1}$, equivalent to 1.5 ± 0.3 and 2.0 ± 0.5 K (3.7 W $m^{-2})^{-1}$, respectively. The latter quantity is equal to the lower bound of the "likely" range for this quantity given by the 2007 IPCC Assessment report. The uncertainty attached to the lower-bound equilibrium sensitivity permits us to state, within the assumptions of this analysis, that the equilibrium sensitivity is greater than 0.31 K (W $m^{-2})^{-1}$, equivalent to 1.16 K (3.7 W $m^{-2})^{-1}$, at the 95% confidence level .

Introduction

Earth's so-called "equilibrium climate sensitivity", the change in global mean near-surface air temperature GMST, T_S, that would ultimately be attained in response to a sustained change of the radiative budget (forcing), ratioed to the forcing, is commonly recognized as a key geophysical property of Earth's climate system and an important index of the susceptibility of the climate system to perturbations in the radiation budget (Hansen, 1984; Meehl et al., 2007). In this definition the global mean near-surface temperature is generally taken as the temperature at 2 m above the ground or ocean surface, in agreement with long-term meteorological practice, and/or a blend of this temperature with sea surface temperature (Brohan et al., 2005; Smith and Reynolds, 2005; Hansen et al., 2010). Global temperature change ΔT_S is generally expressed as anomaly, the spatially averaged change in temperature relative to a specified climatological mean, as anomaly is rather uniform spatially, permitting robust spatial averaging. The forcing is equal to the change in net absorbed irradiance at the top of the atmosphere, TOA (or, alternatively, at the tropopause), due to changes in the absorbed shortwave solar radiation and/or in the emitted long-wave terrestrial radiation that are externally imposed to the climate system, but not including changes in net absorbed irradiance that result from climate system response to the externally imposed change, although this definition leads to some ambiguity, as discussed below.

The magnitude of the equilibrium climate sensitivity depends not only on the Planck response of increased long wave radiation with increased T_s, but also on feedbacks that are consequences of changes in processes that comprise the climate system that occur with changing temperature as the system is attaining a new steady state (commonly denoted "equilibrium") following imposition of a perturbation. Important such changes are changes in atmospheric temperature structure, water vapor, and clouds, and changes in the surface albedo that might result from change in snow and ice cover. The feedbacks thus represent internal processes in the Earth’s climate system. Determining Earth's equilibrium climate sensitivity is a major objective of current climate research, through climate model studies, and empirical approaches through consideration of changes in global temperature and forcing over time during the period of instrumental temperature measurements or from differences in forcing and temperature between the present climate and various paleo climate states, as reviewed by Knutti and Hegerl (2008).

Climate model studies, especially studies with global climate models (GCMs) that represent the major processes comprising the climate system, not only yield estimates of climate sensitivity but also permit determination of the several feedback contributions to this sensitivity. Current models provide similar positive feedback values for atmospheric water vapor and surface albedo but differ considerably for cloud feedback (Bony et al., 2006; Soden and Held, 2006; Webb et al., 2006). These differences, due mainly to differences in the representation of cloud processes, are the principal reason for the spread in climate sensitivity of current GCMs, somewhat more than a factor of 2 (Randall et al., 2007). Despite intense research over the past several decades the range in Earth's climate sensitivity in climate models has hardly decreased and may be expected even to increase as climate models represent increasingly more processes (Maslin and Austin, 2012). . The empirical approach using instrumental temperature data together with estimates of radiative forcing over a specific time period (Gregory et al., 2002; Forster et al., 2007, Gregory and Forster, 2008, Forest et al., 2008; Aldrin et al., 2012; Schwartz, 2012; Otto, 2013) yields substantial uncertainty in inferred climate sensitivity primarily because of large uncertainty in forcing, mainly forcing by tropospheric aerosols. Likewise because of uncertainties in both the forcing and the change in global temperature between the holocene and prior climatic states such as the last glacial maximum, the range of estimates of equilibrium climate sensitivity from paleoclimate studies well exceeds that from climate model studies, especially at the high end of the range (Skinner, 2012; Rohling et al., 2013; Hansen et al., 2013).

The present uncertainty in climate sensitivity has important implications for formulation of policy regarding the amount of additional infrared absorbing gases (so-called "greenhouse gases", GHGs) including CO_2, CH_4 and N_2O that might be emitted consistent with a given acceptable increase in global temperature. As shown by Schwartz et al. (2010), within the range given by the

2007 IPCC Assessment report (Solomon et al., 2007) as the estimated central 66% or more of the probability distribution function for Earth's climate sensitivity, the amount of additional CO_2 equivalent that may be introduced into the atmosphere without committing the planet to an increase in global surface temperature greater than 2 K above preindustrial is uncertain even as to sign. In this context it seems useful to focus on determining a lower bound on climate sensitivity that would allow determination of a firm upper bound to the allowable incremental CO_2 emissions consonant with any maximum acceptable increase in global mean temperature.

In considerations of climate forcing and response it is important to distinguish radiative changes that constitute forcing from those that are part of the climate system response. Consider, for example, a situation in which the solar irradiance incident at the top of the atmosphere were to suddenly exhibit a sustained increase. The forcing would be equal to the planetary co-albedo (complement of albedo) times the change in solar irradiance. In response to this forcing Earth’s climate system would gradually warm, leading to an enhanced terrestrial radiation emitted at the TOA and/or decreased albedo that ultimately would balance the initial increase in absorbed solar radiation. Similarly increases in amounts of GHGs reduce the outgoing terrestrial radiation. The effect is analogous to an increase in the solar radiation, as climate has to warm up to radiate more and thus restore the balance.

From the definition of equilibrium sensitivity given above it is clear that attainment of the new steady-state climate in response to a perturbation occurs over a period of time rather than instantaneously. Increasingly it is becoming recognized that this climate response takes place on multiple time scales. Studies with general circulation models suggest that much of the response, two-thirds to perhaps 80%, occurs on a time scale of a decade or less (Gregory, 2000; Held et al., 2010) following imposition of a forcing. This rapid adjustment, mainly involving the atmosphere, land surfaces, and the upper ocean, results from rapid heat exchange together with limited heat capacity. The part of the adjustment that involves the deep oceans is slow, hundreds of years, because of the huge heat capacity together with relatively weak mixing. During this time period the change in temperature in response to an imposed (positive) forcing is less than the so-called equilibrium response because heat flow from the compartment of the climate system that is closely coupled radiatively to space to the deep ocean diminishes the system response from its "equilibrium" response.

In contrast to the sustained forcing that results from sustained increases in GHGs is the situation forcing by a pulse injection of a material that is removed from the atmosphere over a short period of time as is the situation with cooling forcing by stratospheric aerosols produced by a volcanic eruption. These aerosols, which reflect incident solar radiation thereby cooling the planet, exhibit a time constant for removal from the atmosphere of a year or so. If the incremental GHGs were similarly to disappear within a short period of time, then the previous temperature would be largely restored in a similar way as after a volcanic eruption (Held et al., 2010). However, this is not the case because of the very long atmospheric residence times (multiple decades to centuries) of the so-called long-lived GHGs, LLGHGs, and the fact that these gases are continuously replenished through ongoing anthropogenic emission. Increasing GHGs therefore affects the climate in a similar way as a sustained increase in solar irradiance. Likewise, if for some reason the volcanic aerosols were to remain in the atmosphere indefinitely, the planet would continue to cool until a new, lower steady-state temperature was reached.

The empirical approach is to determine climate sensitivity from known forcings and measured temperature changes. This approach, which relies on the assumption of a cause and effect relation between temperature change and forcing, is attractive but must cover a relatively long period to avoid influences from short-term chaotic weather and climate events. The key required observations for this approach are (i) the net radiative forcing over a period of time F and (ii) the corresponding near-surface temperature change ΔT_S. Because the response of the climate system is not necessarily at steady-state with respect to the imposed forcing it is necessary as well, as discussed below, to know and account for (iii) the planetary radiation imbalance over the time for which the sensitivity is to be inferred from F and ΔT_S. In principle the planetary energy imbalance might be measured from space by satellite-borne radiometers, but at present this approach does not have the required accuracy because of uncertainties in instrument calibration (Loeb et al., 2009) and perhaps as well because of limited sampling. An alternative approach is through measurement of heat accumulation in the Earth's system, some 90% of which is in the oceans and is manifested by increase in ocean temperature; a minor part of the surplus heat is used to warm the atmosphere and to melt ice. As discussed below measurements of ocean temperature with accuracy and geographical coverage sufficient to calculate a change in ocean heat content are available only for the last 40 years or so, limiting the analysis to this period.

The increase in global mean temperature over the past 130 - 160 years is rather well quantified by thermometric measurements. However, the temperature record exhibits fluctuations on a variety of time scales that complicate the analysis. Short-term fluctuations in global temperature are dominated by major volcanic events such as Mount Agung (1963), El Chichon (1982), and Mount Pinatubo (1991), which have affected the global temperature for 1-3 years following the eruption, and by high amplitude ENSO events such as those of 1876-78, 1940-42 and 1997-98. Such short term fluctuations necessitate the use of sufficiently long observational records to reliably determine temperature changes that result from longer term forcing such as the build up of greenhouse gases. Here we focus on the 40-year period 1970-2010. The decision to use this time period is based not only on the need for well examined ocean temperature records but also on the requirement of sufficiently long record for determination of the trend of T_S. The global temperature trend over the period 1970-2010 has been estimated independently by different groups using different analysis methods providing virtually identical results (Brohan et al., 2006, Hansen et al., 2010; Smith and Reynolds, 2005). These results are supported by radiosonde and satellite microwave measurements (after 1979) (Thorne et al., 2010) as well as by recent re-analyses by ECMWF (Simmons et al., 2010).

The forcing required for the empirical method is the total forcing over the period of interest. The radiative forcing by the LLGHGs can be accurately calculated from known changes in their mixing ratios using models that are based on laboratory measurements (Collins et al., 2006; Iacono et al., 2008; Oreopoulos et al., 2012) and evaluated by field measurements (e.g., Turner et al., 2004). However total forcing remains quite uncertain mainly because of uncertainty in forcing by tropospheric aerosols emitted by much the same combustion as has produced incremental CO_2, resulting in large uncertainty in inferred total forcing (Gregory et al., 2002; Forster et al., 2007). Although the radiative effects of aerosols might be estimated from space observations, the accuracy of such determinations is limited especially because of uncertainties in understanding interactions between clouds and aerosols (Stevens and Schwartz, 2012). As emission of aerosols and precursor gases is related to the use of fossil energy (mainly coal), in view of the continued increase in combustion in the period 1970-2010 (Boden et al., 2010; IEA, 2011) it seems unlikely that there has been a decrease in aerosol cooling forcing over this period. This supposition is reflected also in model based estimates of aerosol forcing; for example the estimate of the increase in total aerosol forcing (direct plus indirect) over the period 1970-2005 in the Representative Concentration Pathways data set (RCP; Meinshausen et al., 2011; http://www.pik-potsdam.de/~mmalte/rcps/) that is widely used in climate modeling studies of the twentieth century is highly correlated with the increase in LLGHG forcing (proportionality coefficient -0.24; $r^2 = 0.94$). In the present analysis we restrict consideration of forcing only to that due to the increase in LLGHG concentrations. As any increase in aerosol cooling forcing would decrease the net forcing from that due to increases in LLGHG concentrations, we consider the climate sensitivity determined using only the LLGHG forcing to be a lower bound on the actual value. In that respect this study differs from others (Gregory and Forster, 2008; Schwartz, 2012) that have provided estimates of climate sensitivity based on estimates of total forcing, rather than just LLGHG forcing. Forcings by volcanic aerosols are not considered because their short duration; forcing by solar variability is likewise not considered, because of its small magnitude and periodic nature.

Two measures of climate sensitivity are examined, the *equilibrium sensitivity*, as defined above, and the proportionality between the increase in T_S and imposed forcing that is achieved on decadal time scales that has been examined by several investigators (Dufresne and Bony, 2008; Gregory and Forster, 2008; Held et al., 2010; Padilla et al., 2011; Schwartz, 2012) and has been denoted (Held et al., 2010; Padilla et al., 2011; Schwartz, 2012) as the *transient climate sensitivity*. As this transient sensitivity does not account for the planetary energy imbalance, it is less than the equilibrium sensitivity and is thus a further and less restrictive lower bound on the equilibrium climate sensitivity.

In distinguishing the transient and equilibrium climate sensitivities, it would seem that for many purposes the transient climate sensitivity may be a more useful quantity than the equilibrium sensitivity. As the major fraction of climate system response to a sustained perturbation is likely reached within a decade of so of the onset of the forcing, and as the remainder of the response takes place only over multiple centuries the transient sensitivity is pertinent to the change in global temperature that would expected on societally relevant time scales. Further, as the atmospheric burden of incremental LLGHGs subsequent to and attributable to a given set of emissions would be expected to decay on the time scale of multiple decades to centuries, depending on the substance, the long term committed temperature increase from a given emitted amount of these gases would decrease over much the same time period as the remaining temperature increase between the shorter term response characterized by the transient sensitivity and the longer term ("recalcitrant"; Held et al., 2012) response characterized by the equilibrium sensitivity. For this reason, as well as the practical reason of being able to infer the transient sensitivity from observations over a few decades, we focus attention on both the transient and equilibrium sensitivities. As discussed below (also, Schwartz, 2012) the two quantities are related by the planetary heating rate, allowing the equilibrium sensitivity to be inferred from the transient sensitivity.

Here we use observational data (temperature change, planetary heating rate) and model estimates of forcing by incremental LLGHGs over the period 1970-2010 to adduce a firm lower bound to Earth's transient and equilibrium climate sensitivities that can serve as a confident basis for minimum actions necessary to avert a given committed increase in global temperature. Although it must be recognized that planning based on such a lower limit sensitivity may not result in emissions limitations that are sufficient to confidently avert such a temperature increase, the minimum sensitivity has the value of providing a firm floor for such emissions reductions. We thus focus on the lower limit sensitivity, rather than any specific emissions strategies required to meet a particular maximum allowable increase in global temperature.

Commonly Earth's equilibrium sensitivity is reported as the temperature change $\Delta T_{2\times,eq}$ that would result from a sustained forcing $F_{2\times}$ equal to that due to a doubling of atmospheric CO_2, taken as approximately 3.7 W m^{-2} (Meehl et al., 2007; Myhre et al., 1998). Thus an equilibrium climate sensitivity S_{eq} of 1 K (W $m^{-2})^{-1}$ would be equivalent to the more familiar equilibrium doubling temperature $\Delta T_{2\times,eq}$ of 3.7 K. To facilitate comparison we therefore also present sensitivities in the unit K (3.7 W $m^{-2})^{-1}$.

Theoretical framework

To good approximation Earth's energy budget is given by

$$\frac{dH}{dt} \equiv N = Q - E \qquad (1)$$

where H is a measure of the amount of heat content of the Earth’s climate system (atmosphere, ocean, land areas, and the cryosphere)

N is the net change in planetary heat content with time t

Q is the absorbed short wave irradiance at the TOA, and

E is the emitted long wave irradiance at the TOA.

The two fluxes Q and E are approximately the same magnitude, ca 240 W m^{-2}, with the difference N being much smaller, 1 W m^{-2} or less.

If a time-dependent perturbation, a so-called “forcing”, $F(t)$, is applied to a system initially at steady state, inducing a change in the global heat balance, the energy budget becomes:

$$N(t) = F(t) + Q(t) - E(t) \qquad (2)$$

In response to the perturbation the global mean surface temperature T_s will change, inducing a response in the radiation budget. This response may be expressed in terms of the change in T_s, ΔT_s, as

$$N(t) = F(t) + Q_0 - E_0 - \lambda \Delta T_s(t) + \text{ higher order terms} \qquad (3)$$

where $\lambda \equiv -\partial(Q-E)/\partial T_s$ is denoted the climate response coefficient; the minus sign is used in order to let λ be a positive quantity; the partial derivatives denote response of the radiation to change in surface temperature, i.e., excluding the forcing itself. The response coefficient λ (units W m^{-2} K^{-1}) describes the climate system response to the forcing. In principle the higher order terms in Eq (3) would account for different climate responses to forcings that are different in nature (e.g., solar, greenhouse gas, aerosol) and/or spatial distribution, resulting in different spatial or seasonal patterns of temperature change for the same change in global mean temperature. Climate model studies indicate that the differences in the global mean sensitivity for different kinds of forcings are fairly small, typically less than 20% (e.g. Hansen et al., 1997; Boer and Yu, 2002; Joshi et al., 2003; Kloster et al., 2010), supporting the climate sensitivity concept. The higher order terms would also reflect any change in sensitivity with global mean temperature, i.e., second-derivative terms. Such effects are neglected in the present analysis.

For a system initially at steady state prior to the imposition of the forcing, $Q_0 = E_0$, and hence

$$N(t) = F(t) - \lambda \Delta T_{\mathrm{s}}(t) + \text{ higher order terms} \qquad (4)$$

If the forcing were maintained constant until the system reached a new steady state ($t = \infty$), then

$$\Delta T_{\mathrm{s}}(\infty) = \lambda^{-1} F , \qquad (5)$$

from which the identification can be made between the equilibrium sensitivity, S_{eq}, the ultimately achieved ratio of temperature change to forcing, and the climate response coefficient λ

$$S_{\mathrm{eq}} = \lambda^{-1} . \qquad (6)$$

Equation (6) allows the time-dependent response of temperature to be expressed in terms of the equilibrium sensitivity as

$$\Delta T_{\mathrm{s}}(t) = S_{\mathrm{eq}} \left[F(t) - N(t) \right]. \qquad (7)$$

Equation (6) explicitly shows the effect of the global heating rate in diminishing the increase in T_{s} from its "equilibrium" value.

In general, and more specifically with respect to the response of Earth's climate system to the perturbation of forcing over the industrial era, the climate system is not in steady state because of the high thermal inertia of the system that is due to the huge heat capacity of the oceans and resultant large time constant for reaching steady state. Hence N is not equal to zero but is expected to be positive; less than, but of comparable magnitude to, the imposed forcing $F(t)$. $N(t)$ is thus a measure of the imbalance in the radiation as the global temperature has not yet fully adjusted to imposed forcing. That this is the case for Earth's climate system at present can be seen from the ongoing warming of the world ocean, as observed in measurements of the increase in heat content of the global ocean, as examined in Sec. 3.

Equation (7) serves as the basis for observational determination of the equilibrium climate sensitivity as

$$S_{eq} = \frac{\Delta T_s(t)}{F(t) - N(t)} = \left(S_{tr}^{-1} - \frac{N(t)}{\Delta T_s(t)} \right)^{-1} \quad (8)$$

The transient climate sensitivity S_{tr} is obtained as the change of the observed global temperature over a period of time relative to the change in forcing over that period,

$$S_{tr} = \frac{d\Delta T_s(t)}{dF(t)}, \quad (9)$$

where the change in T_s over a period of time is inferred from observations of the global temperature record, and where the forcing is calculated from changes in atmospheric composition that are externally imposed on the climate system (as distinguished from changes in water vapor that are part of the climate system response). Specifically in the present study we restrict consideration of forcing to that arising from changes in mixing ratios of the LLGHGs, mainly CO_2, CH_4, N_2O and chlorofluorocarbons F11 and F12. The values of $N(t)$ and $\Delta T_s(t)$ to be employed in Eq (8) are values of these quantities over the time of determination of S_{tr}. Here it is important that $\Delta T_s(t)$ represent the change in global temperature relative to a steady-state (unforced) situation that is responsible for the climate system response term in Eq (3). For this analysis we use measurements of T_s relative to the beginning of the twentieth century, which we take as representative of the planetary temperature prior to any substantial response to GHG forcing.

To determine the planetary heating rate N we use measurements of ocean heat content. As the ocean is the principal means of storing heat in the climate system, at least on the multi-century to millennial time scale, we obtain a first approximation to N from the time derivative of ocean heat content, to which we add corrections for other heat sinks.

Analysis and results

Forcing, temperature anomaly change, and transient sensitivity

Several GHG forcing data sets were examined (Figure 1) to span the time range of interest and to assess the spread in current estimates. It should be emphasized that these forcings and indeed all estimates of forcings are based on globally averaged radiation transfer calculations for perturbed atmospheric composition rather than direct measurement, although the radiation transfer calculations are strongly supported by measurements (e.g., Turner et al., 2004). The NOAA Annual Greenhouse Gas Index (http://www.esrl.noaa.gov/gmd/aggi/) presents forcing only from

1979 to the present. For the times in which the two data sets overlap this forcing closely matches that of the RCP data set, which ends in 2005 but extends back in time to 1860. Because of the close match between the two data sets the two sets are combined into a single record for the present analysis, denoted here as the "blended" forcing. The second independent forcing data set examined here, that of the NASA GISS group (Hansen et al., 2007; http://data.giss.nasa.gov/modelforce) increases at an appreciably greater rate throughout the entire period.

Comparison of GHG forcing and temperature anomaly over the period of instrumental temperature records shows good qualitative correlation, Figure 2; the ratio of the scales of the vertical axes in the figure (0.314 K (W $m^{-2})^{-1}$ was determined by the slope of a least-squares fit of temperature anomaly to forcing which exhibited a correlation $r^2 = 0.77$; similar slope (0.258 K (W $m^{-2})^{-1}$) and correlation coefficient (0.82) are found with the GISS forcing and temperature record (Hansen et al., 2010; http://data.giss.nasa.gov/gistemp/). This correlation of temperature change with GHG forcing contributes to the attribution of the warming over this period to the increase in GHG forcing that is the premise of the present analysis. The quantitative examination of the correlation leading to the present estimates of climate sensitivity is limited to the time period subsequent to 1970 for which T_s is more or less monotonically and systematically increasing and for which globally representative ocean heat content data are available.

Detailed comparison of the two forcing data sets for the time period 1970-2010, Figure 3, again shows the somewhat greater GHG forcing in the GISS data set relative to the blended RCP-NOAA data set, 0.31 W m^{-2} out of a total increase over this time period 1.62 W m^{-2}. In the analysis presented here we use the average of the two forcings and take the difference between the average and either of the two forcings (± 9.6%) as a measure of uncertainty. This uncertainty is virtually identical with the ± 10% uncertainty (5 – 95% of the probability distribution function, equivalent to ± 1.64 σ, i.e., 1-σ uncertainty 6.1%) that is given by the 2007 IPCC Assessment Report (Forster et al., 2007) and by earlier IPCC Assessments for forcing by the LLGHGs, but we consider this difference more of a 1-σ uncertainty as it is based on the actual difference between the two estimates and treat it as thus. Unless otherwise indicated, all uncertainties presented here are 1-σ estimates.

An alternative approach to estimating the uncertainty associated with forcing by LLGHGs is through examination of the spread of forcings in current GCMs. Recently Andrews et al. (2012) compared CO_2 forcings and climate response of 15 atmosphere-ocean general circulation models (GCMs) that participated in the Coupled Model Intercomparison Project CMIP-5. Forcing and temperature response coefficient were inferred from the output of the model runs respectively as intercept and slope of a graph of net top-of-atmosphere energy flux versus global mean temperature anomaly subsequent to a step-function quadrupling of atmospheric CO_2. (Because the model experiments examined response to a quadrupling of CO_2, rather than a doubling, the intercept had to be divided by 2 to obtain the forcing pertinent to doubled CO_2). The forcing is interpreted as an "adjusted forcing" that includes rapid adjustments, mainly of atmospheric structure, that modify the TOA radiative flux on time scales shorter than a year or so. A key finding of the Andrews et al. study was the spread of values of forcing exhibited by the different GCMs, 16%, 1-σ. The spread in forcing is a consequence of differing treatments of the radiation transfer in the several models as well as different treatments of clouds that interact with radiation. As the forcing inferred from the analysis of Andrews et al. is an adjusted forcing, it appropriately reflects differences among the models in rapid ($\lesssim$ 1 yr) response of atmospheric structure to the imposed forcing. This spread in forcings inferred from the climate model runs is substantially greater than the uncertainty specified in the IPCC Report. It would seem that it is this uncertainty that should be combined (in quadrature) with the uncertainty in ΔT_s over a time period of interest to obtain an accurate measure of the uncertainty in observationally derived minimum transient climate sensitivity.

As noted earlier we calculate a minimum transient sensitivity that is based only on forcing by the LLGHGs, neglecting other contributions to climate forcing over this time period. For the reasons given above we consider the change in forcing over the period 1970-2010 to be dominated by the increase in LLGHG forcing (of which about 60% is due to increases in CO_2, with the balance due to increases in other LLGHGs; Meinshausen et al., 2011). Principal other contributions are short wave forcing by anthropogenic and natural (volcanic) aerosols, long wave forcing by tropospheric ozone, and variability in solar irradiance, of which the shortwave aerosol forcing exhibits the greatest magnitude and uncertainty. To assess the magnitude of forcings by agents other than the LLGHGs we also show in Figure 3 the difference between the total forcing and the LLGHG forcing for the RCP and GISS forcing datasets. Most prominent in the figure are the (negative) forcings from stratospheric aerosols produced by eruptive volcanos (Fuego, 1974; El Chichon, 1982; Pinatubo, 1991), but these forcings disappear on a time scale of two years or so and thus contribute little to the long term trend, especially as there has been little volcanic activity subsequent to the 1991 Pinatubo eruption through 2010 (Sato et al., 1993, as updated; Gao et al., 2008; Solomon et al., 2011; Bourassa et al., 2012). The balance of the non-GHG forcing is due mainly to tropospheric aerosols. The two forcing data sets suggest that this forcing is rather small, less than 0.5 W m^{-2} (magnitude) and, more importantly in the present context, does not exhibit substantial trend over the period. A cautionary note about these estimates is that the magnitude of the forcing in these two estimates is well less than the uncertainty associated with present estimates of year-2005 aerosol forcing, for which the 2007 IPCC Assessment report (Forster et al.

2007) gives -0.5 [-0.1, -09.] W m^{-2} for the direct effect and 0.7 [-0.3, -1.8] W m^{-2} for the indirect effect, where the square brackets indicate the 5-95% confidence range.

A graph of ΔT_s vs LLGHG forcing evaluated with the average of the GISS and blended RCP-NOAA data sets, Figure 4, exhibits a correlation coefficient $r^2 = 0.80$ indicative of a fairly robust correlation over this period and a slope of 0.39 K (W $m^{-2})^{-1}$ with standard error 0.03 K (W $m^{-2})^{-1}$. This number is given in Table 1. We also examined the sensitivity of slope to start date of the regression over the years 1960-80, finding a standard deviation of the slope so obtained to be 0.03 K (W $m^{-2})^{-1}$. However because of the difference in forcing between the two data sets shown in Figure 4, we consider the uncertainty associated with the slope to be an underestimate of the uncertainty associated with transient sensitivity ; we therefore combine the further uncertainty in forcing (taken as 16%, 1-σ, as discussed above) with that associated with the slope to yield an uncertainty (1-σ) of 0.07 K (W $m^{-2})^{-1}$. According to Eq (9) the slope of this graph would correspond to the transient climate sensitivity S_{tr} over this time period if the forcing employed in the graph were the total forcing; as the forcing is for LLGHGs only, and as the change in LLGHG forcing is likely to be fairly close to or perhaps slightly greater than the change in total forcing, we consider the transient sensitivity obtained in this way a fairly confident estimate of the actual value that characterizes the normalized transient response of Earth's climate system to a forcing, although a somewhat greater value cannot be ruled out, given the uncertainty in aerosol forcing. We thus consider this value to be a fairly robust best-estimate lower bound to Earth's transient climate sensitivity. Finally, when the uncertainty on this estimate is taken into account we obtain, as the lower bound of the 5-95% confidence range (1.64 σ) 0.28 K (W $m^{-2})^{-1}$, for the probability distribution function for the quantity taken as normally distributed. We also present in Table 1 the value of S_{tr} so obtained in the unit K (3.7 W $m^{-2})^{-1}$, the 3.7 W m^{-2} being the forcing commonly given (Myhre et al., 1998) for doubled CO_2, $F_{2\times}$, to obtain a measure of best-estimate lower-bound sensitivity $S_{tr} = 1.46 \pm 0.26$ K (3.7 W $m^{-2})^{-1}$ that can be compared with the CO_2 doubling temperature commonly used to express Earth's climate sensitivity. This quantity is well below the range of current estimates for the equilibrium doubling temperature, 2 – 4.5 K. To some extent the lower value obtained in this way is due to the quantity being a measure of transient, not equilibrium, sensitivity, and to some extent because it is based on forcing by LLGHGs only.

Table 1. Calculation of lower bound transient and equilibrium sensitivities.

Quantity	Unit	Best estimate	1-σ Uncertainty	Lower 5% Bound
ΔF (1970 - 2010)	W m^{-2}	1.465	0.234	
S_{tr}	K (W m^{-2})$^{-1}$	0.394	0.071	0.278
S_{tr}	K (3.7 W m^{-2})$^{-1}$	1.460	0.262	1.031
N	W m^{-2}	0.374	0.032	
ΔT_s (1900-1990)	K	0.529	0.100	
S_{eq}	K (W m^{-2})$^{-1}$	0.545	0.142	0.312
S_{eq}	K (3.7 W m^{-2})$^{-1}$	2.023	0.528	1.157

LLGHG forcing over period 1970-2010 ΔF is based on mean of GISS and blended RCP-NOAA forcing data sets; uncertainty in forcing is taken as ± 16% as discussed in text. Column 3 presents values for forcing by LLGHGs only and thus yields a best estimate for lower bound transient and equilibrium sensitivity. Uncertainty in S_{tr} reflects uncertainties in ΔF and $d\Delta T_s/d\Delta F$. Heating rate N and associated uncertainty are from Table 2. Time range for ΔT_s is for middle of time period examined relative to assumed steady state at beginning of twentieth century. Last column shows lower bounds of the 5-95% uncertainty range, evaluated as the best-estimate value of the lower bound minus 1.64 times the 1-σ uncertainty for the probability distribution function for the quantity taken as normally distributed. Values of S_{tr} and S_{eq} expressed in the unit K (3.7 W m^{-2})$^{-1}$ are shown to permit comparison with commonly reported CO_2 doubling temperature $\Delta T_{2\times}$.

Planetary heating rate and equilibrium sensitivity

As noted above, the planetary heating rate must be subtracted from the forcing in order to infer the equilibrium climate sensitivity from observations. Although N cannot be determined from satellite measurements it can, as discussed above, be estimated from the rate of heat accumulation in the oceans. As the principal contribution to planetary heat uptake in response to forcing is heating of the global ocean, much effort has been made in recent years to archive and analyze measurements of ocean temperature, permitting determination of heat content anomaly (referenced to a given time period) as the volume integral of local heat content anomaly evaluated as temperature anomaly times heat capacity. For recent reviews see Palmer et al. (2009), Church et al., (2011), and Lyman (2012). Recently Levitus et al. (2012) presented a new assessment of ocean heat accumulation from the surface to 2000 m that we make use of in this article. Although the data presented by Levitus et al. cover the period from 1955 to 2011 (Figure 5), prior to 1970 the observational network is very sparse. From around 1970 onwards a systematic, approximately linear increase in heat accumulation is noted with rate $0.48 \pm 0.02 \times 10^{22}$ J yr^{-1}; this corresponds to an average heating rate, expressed per area of the planet, of 0.30 ± 0.01 W m^{-2}.

Other key sinks for heat taken up by the planet in response to forcing are heating of the ocean below 2000 m, heating of the atmosphere and the upper land surface, and melting of sea ice, sea-shelf ice, and ice in glaciers and small ice caps. Levitus et al., were unable to present a value for ocean heat uptake below 2000 (the average depth of the oceans is ca 3800 m), but it would seem that this additional heat uptake can be no more than about 20% of the amount above 2000 m (see Figure 2 of Levitus et al.). We thus augment the ocean heating rate to 2000 m by 10% and place a 10% uncertainty on the estimate. The magnitudes of other heat sinks were examined by Hansen et al (2011) whose estimates, summarized in Table 2, constitute an additional 14% relative to the ocean heating rate. The total heating rate of the planet for the years 1970-2010 is thus estimated as 0.37 ± 0.03 W m^{-2}. This heating rate agrees closely with that recently given by Otto et al. (2013; supplementary information) 0.35 ± 0.08 W m^{-2} (uncertainty adjusted from original to denote 1 σ value).

Table 2. Contributions to planetary heating rate.

Component	Heating rate W m^{-2}	Uncertainty W m^{-2}	Start Yr	End Yr
Atmosphere	0.0057	0.0003	1980	2007
Land	0.0187	0.0006	1980	2006
Sea ice melt	0.0072	0.0005	1981	2007
Ice shelf melt	0.0022	0.00003	1982	2007
Ice sheet melt Greenland, Antarctica	0.0049	0.0002	1982	2006
Glaciers, small ice caps	0.0077	0.0002	1982	2007
Total non-ocean	0.0464	0.0009		
Ocean to 2000 m	0.298	0.012	1970	2008
Ocean below 2000 m	0.030	0.030	1970	2008
Total ocean	0.327	0.032	1970	2008
Total	0.374	0.032		

Non-ocean components of Earth's energy imbalance are based on Hansen et al. (2011). The rate of ocean heating, from Figure 5, is based on Levitus et al. (2012).

Comparison of this planetary heating rate to the increased radiative forcing by incremental LLGHGs during the same period, 1.46 ± 0.16 W m^{-2}, indicates that the heating of the planet decreases the effective forcing over this period by about 25%. This simple calculation would suggest that the equilibrium sensitivity should be about $(0.75^{-1} -1) = 33\%$ greater than the transient sensitivity calculated for this period, or about 0.53 W m^{-2}. A more explicit calculation by Eq (7) yields the result $S_{eq} = 0.55 \pm 0.14$ K (W $m^{-2})^{-1}$ equivalent to 2.0 ± 0.5 K (3.7 W $m^{-2})^{-1}$. This value, which coincides with the low end of the range for equilibrium climate sensitivity expressed as CO_2 doubling temperature as given by the IPCC Assessment (Solomon et al., 2007), is an independent robust estimate of this lower limit equilibrium sensitivity.

Finally, we take into account the uncertainties in the values of S_{tr} and S_{eq} obtained in this way, which we express as the value below which the actual value of the quantity is estimated as having a probability of 5%, evaluated by multiplying the 1-σ uncertainty by 1.64, and subtracting from the central value. In this way we obtain what we denote as lower bounds for S_{tr} and S_{eq} of 0.28 and 0.31 K (W $m^{-2})^{-1}$ equivalent to 1.03 and 1.16 K (3.7 W $m^{-2})^{-1}$, respectively. These lower bounds are well below the low end of the range for equilibrium climate sensitivity given by the IPCC 2007 Assessment, a consequence of the uncertainties in the estimated sensitivities, 18% and 26% (1-σ) for the transient and equilibrium sensitivities, respectively. Examination of the sources of uncertainty in these quantities shows that it arises mainly from the uncertainty in the forcing by LLGHGs, which we have taken as 16%, 1-σ. As noted above, this uncertainty is substantially greater than that given by IPCC Assessments, 6.1% (1-σ), but for the reasons stated above we feel that lower uncertainty estimate cannot be justified.

The transient and equilibrium sensitivities determined here are based on the assumption, surely incorrect, that forcing by LLGHGs is the sole secular forcing change over the period 1970-2010. The principal other forcing is that due to tropospheric aerosols, and as noted above this forcing is highly uncertain. It would seem, however, that any incremental aerosol forcing over this period is almost certainly well less (in magnitude) than the incremental LLGHG forcing. Because the aggregate of other forcings, including tropospheric aerosol forcing, is almost certainly negative (i.e., exerting a cooling influence), Figure 3, the sensitivities based only on incremental LLGHG forcings are almost certainly lower bounds to the actual sensitivities characterizing Earth's climate system.

Discussion

Earth's equilibrium climate sensitivity is a key geophysical property of the Earth's climate system, the ratio of the annually averaged change in global mean near-surface temperature T_s to radiative forcing, indefinitely maintained, once the climate system has reached a new steady state. Earth's transient climate sensitivity is the ratio of the change in surface temperature to forcing, but without the requirement that a new steady state has been reached. It is less than the equilibrium sensitivity because the rate of heating of the planet serves as a heat sink in addition to radiation at the top of the atmosphere. The two sensitivities are related by this heating rate, Eq (8). We have provided best estimates for the lower bounds for both the transient and the equilibrium climate sensitivity, Table 1.

Determining equilibrium climate sensitivity from empirical data requires accurate information on near-surface temperature, the net heat flux into the Earth's system, and the forcing at the top of the atmosphere. We claim that reliable such data exist for the period 1970-2010 with the exception of accurate forcing data, mainly because of uncertainty in forcing by tropospheric aerosols.

We estimate the uncertainty in the increase in global temperature over the 40-year period examined here to be less than 0.05° C. This is supported by the close agreement of available data sets including radiosonde data and microwave measurements from the lower troposphere (Thorne et al., 2010). We note that temperature trend over land is ca. 3 times larger than over oceans and it cannot be excluded that land temperatures in some regions are influenced by factors other than those related to direct or indirect effects of the LLGHGs, such as excessive agriculture or forestry changes.

Because of uncertainty in the forcing data it is not possible to determine a specific value for climate sensitivity. However, by considering only the forcing by LLGHGs it is possible to determine robust and useful lower limits of the transient and equilibrium sensitivities. We consider the lower-limit estimates obtained in this way to be robust on several grounds. From the perspective of emissions it seems highly unlikely that the production of tropospheric aerosols associated with fossil fuel combustion has decreased between 1970 and 2010.

The change in aerosol forcing over the period 1970-2010 is very difficult to assess as in this period there was a reduction of SO_2 emission in North America and Europe but an increase in China and India. According to IEA (Key World Energy Statistics, 2011) the burning of coal, the main source of sulphate aerosol, has in this time (1973-2009) risen at about the same rate (2.2% yr^{-1}) as the total forcing contribution by LLGHGs (2.3% yr^{-1}). Likewise the production of secondary organic aerosols, the second major component of anthropogenic aerosols (Zhang et al., 2007), would be expected to scale with fossil fuel combustion, as the photochemistry responsible for production of these aerosols is driven mainly by emissions of nitrogen oxides associated with fossil fuel combustion (DeGouw and Jimenez, 2009). Another complicating factor is that some aerosol substances, particularly black carbon, contribute a warming forcing. Emission of black carbon has been increasing in recent decades, especially in rapidly developing nations (Bond et al., 2007). If, as suggested (e.g., Ramanathan and Carmichael, 2008) this black carbon contributes substantially to climate forcing, then the increase in forcing over the 1970-2010 period would be greater than that due to the incremental greenhouse gases alone, and hence the actual climate sensitivities would be less than the minimum values we report.

A key means of assessing the change in aerosol forcing over time is through satellite measurements. In particular the AVHRR (Advanced Very High Resolution Radiometer) instrument has been in operation throughout much of the time period and might be expected to provide a homogeneous set of measurements (Ignatov and Stowe, 2002) despite the limited wavelength coverage (two bands in the shortwave), restriction to measurements over oceans, concerns over calibration stability, concerns over contamination from clouds, glint and whitecaps, and sensitivity of retrieved AOD to assumptions about real and imaginary components of refractive index and phase function (Wagener et al., 1997; Mishchenko et al., 1999; 2012). Examination of the loading of anthropogenic aerosol is limited to years in which volcanic contribution to AOD is minimal. From examination of the time series of AOD from AVHRR retrievals Mishchenko et al. (2007) reported a significant systematic decrease in AOD over the years 1994-2005, a period minimally influenced by volcanic aerosols. Such a decrease would call into question the assumption made here that aerosol forcing is not decreasing over the time period employed here (1970-2010) of the determination of minimum climate sensitivity. However subsequently these investigators (Mishchenko et al., 2012) reported that the retrieved AOD is highly sensitive to assumed imaginary component of refractive index such that within reasonable assumptions on this quantity there is essentially no change in global and hemispheric AOD between 1985 and 2006, supporting the assumption of the present study.

Although available only for a shorter time record the MODIS (Moderate Resolution Imaging Spectroradiometer) and MISR (Multi-angle Imaging Spectroradiometer) satellite instruments are less subject to the interferences and biases associated with retrievals of AOD by AVHRR. Remer et al (2008; their figure 5) found no discernible trend in global over-ocean AOD as determined by MODIS on both Terra and Aqua platforms over the period 2002 - 2006. Subsequently Zhang and Reid (2010), examining mid-visible over-ocean AOD as determined from the ten-year (2000–2009) Terra MODIS and MISR aerosol products and 7 years of Aqua MODIS, found a statistically negligible global trend in AOD of 0 ± 0.003 per decade. A similar conclusion was reached by Stevens and Schwartz (2012) based on lack of trend of AOD from MISR measurements of and lack of trend of upwelling shortwave irradiance in cloud-free regions as measured from satellite by CERES (Clouds and Earth's Radiant Energy System).

Taken as a whole, the satellite observations lend strong support to the assumption employed in our analysis of little or no decrease in loading of anthropogenic aerosols over this time period and in turn the conclusion that the climate sensitivities determined under that assumption are minimum values. In fact, the small change in AOD indicated in those studies suggests that the actual transient and equilibrium sensitivities may be fairly close to the minimum values that we report in Table 1.

Estimating equilibrium climate sensitivity from transient sensitivity requires information on the global radiative imbalance (planetary heating rate). Although in principle this quantity might be estimated by satellite measurements, current measurements lack the required accuracy or precision. Consequently we use estimates of the accumulation of heat in the Earth climate system determined mainly from measurements of ocean temperature as a function of time, with the heating rate determined as the time derivative. It is possible to do this for the period 1970-2010 but hardly for any earlier period. The Levitus assessment of heating rate, is lower than other current estimates (Lyman, 2012 and references therein) but is more comprehensive and for that reason more relevant for this study. The heat accumulation in the ocean below 2000 m is poorly known and we have expressed this with a significant error bar. However, as the heating rate below 2000 m is certainly much smaller than that above 2000 m, the uncertainty in this heating rate is of little consequence.

We have not intended here to determine a best estimate or an upper bound to climate sensitivity, both of which would require reliable data on aerosol forcing, as noted by Gregory et al (2002), who were unable to determine an upper bound to equilibrium sensitivity for the same reason. Schwartz (2012) presented a similar analysis for a range of forcings employed in recent modeling studies and showed that this range of forcings resulted in a wide range for equilibrium sensitivity, 0.31 ± 0.02 to 1.32 ± 0.31 K (W m^{-2})$^{-1}$. Here the more limited time span and the small change in aerosol forcing over this period, together with improved estimates of planetary heating rate, permit determination of a fairly robust lower-bound estimate of climate sensitivity.

The quantity that we have denoted as the lower bound minimum equilibrium sensitivity, that is, our best estimate of the minimum sensitivity minus 1.64 times the 1-σ uncertainty associated with this best estimate, corresponding to 95% of the probability distribution function, PDF, (taken as normally distributed) of the minimum sensitivity, 0.31 K (W m^{-2})$^{-1}$ or 1.15 K (3.7 W m^{-2})$^{-1}$ (Table 1) is essentially equal to the no-feedback Planck sensitivity of Earth's climate system. From this we conclude that it is "very likely" (in the sense used by the IPCC Fourth Assessment Report, 2007) that net climate feedback is positive relative to the Planck sensitivity, or equivalently that it is "very unlikely" that this net feedback is negative. This lower bound is also essentially equal to the "likely" (84% of the PDF) lower bound of climate sensitivity given by the 2007 IPCC Assessment Report. The present observationally based analysis would thus seem to yield a firmer estimate of the lower bound of climate sensitivity than that given by the 2007 IPCC Assessment.

Although the transient climate sensitivity examined here is somewhat different from the so-called "transient climate response" of global climate models, evaluated as the increase in global temperature in a climate model run during which CO_2 mixing ratio is increased at a compound rate of 1% yr^{-1} at the time (70 years) at which CO_2 mixing ratio is twice its initial value, it seems useful to compare these quantities as both quantities are a measure of climate response to a ramped forcing. It has been suggested (e.g., Meinshausen et al., 2009) that the transient climate response may in fact be a more useful quantity for policymaking than the equilibrium climate sensitivity because of the long time (centuries) associated with reaching a new steady state. The transient climate response of the climate models examined in the IPCC Fourth Assessment (Randall et al., 2007) varies between 1.2 K and 2.6 K, with a mean value of 1.9 K. These values may be compared to the best-estimate minimum value of S_{tr} obtained here, 0.39 K (W $m^{-2})^{-1}$ or 1.46 K (3.7 W $m^{-2})^{-1}$ (Table 1), with a 5% lower bound of 0.28 K (W $m^{-2})^{-1}$ or 1.03 K (3.7 W $m^{-2})^{-1}$. The minimum transient climate sensitivity determined here is thus at the low end of the range of transient climate response exhibited by the climate models and is thus consistent with those results.

A puzzling factor, noted above, is the modest warming since the end of the 19th century that amounts only to some 0.8 K. The forcing of the greenhouse gases so far amounts to 2.8 W m^{-2}. If the observed warming were due only to greenhouse gas forcing, then we would arrive at a very low climate sensitivity of 0.31 K (W $m^{-2})^{-1}$ or 1.16 K (3.7 W $m^{-2})^{-1}$, Fig. 2. Either there was a compensating increasing trend in negative (cooling) forcing over this period due to increasing aerosols or, in the alternative extreme, the climate sensitivity is actually that low and, over the period 1970-2010 there was no increase in the cooling aerosol forcing.

If we were to assume an incremental negative (cooling) aerosol forcing over the period 1970-2010 of -0.5 W m^{-2} then the resulting value of the transient sensitivity would be S_{tr} = 0.60 K (W $m^{-2})^{-1}$ substantially greater than the lower bound sensitivity given in Table 1. The corresponding equilibrium sensitivity is 1.07 K (W $m^{-2})^{-1}$ [3.97 K (3.7 W $m^{-2})^{-1}$], a value more or less in agreement with some climate model results. However, as noted above, there is no support in observations for such an increase in the magnitude of aerosol forcing~~.~~

The parameters used in these estimates must be considered open to further refinement. The forcing of the enhanced greenhouse gases, which is probably the most reliable, is expected to be correct within some 16%, 1-σ. Other contributing forcings, in particular those due to different kinds of aerosols, are not very well known. As discussed above, the net aerosol contribution in the period 1970 – 2010 was probably rather small, but a modest increase cannot be excluded. Nor for that matter is it possible to exclude a minor reduction in the overall contribution from cooling aerosols in this period, but this seems less likely. Based on the foregoing considerations we feel rather confident that the values of transient and equilibrium climate sensitivity determined here constitute robust lower bounds.

The equivalent CO_2 mixing ratio today (for present forcing by LLGHG of 2.8 Wm^{-2}) corresponds to ca 475 ppm CO_2. An equivalent CO_2 mixing ratio of 560 ppm, equal to a doubling of the pre-industrial value, is expected to be reached in some 30 years, or around 2040. If the transient sensitivity is equal to the best estimate lower bound value determined here, 0.39 K $(W\ m^{-2})^{-1}$ and if aerosol forcing remains roughly constant at its present value, the further increase in greenhouse gas forcing would result in a further temperature increase over this time of ca 0.34 K in addition to the ca 0.8 K warming that has occurred already, at an average rate of some 0.11 K per decade. As this temperature increase is based on the lower bound transient sensitivity, it is a lower bound to the actual increase in temperature that would be expected.

Studies with coupled atmosphere-ocean climate models show that transient response to a step-function forcing that is reached within a decade or so of imposition of a forcing comprises the great majority (75% or more) of the total response. For this reason we suggest that transient climate sensitivity is more useful than equilibrium sensitivity for policy purposes such as developing strategies to limit the increase of global temperature to a particular value. Additionally transient sensitivity can more readily and more confidently be determined from observations. Consequently we recommend that increased attention be directed to determination of transient sensitivity in models and observations.

A critical issue is whether a time period of 40 years is sufficient to infer a climate sensitivity given fluctuations in global mean temperature in observations and coupled AOGCM calculations on such time scales. In this respect it is reassuring that an alternative estimate of S_{tr} obtained for the whole period 1860-2012 is very close to the minimum value, including the two standard deviations, obtained in this analysis.

The question also arises whether a measure of global temperature change obtained using only ocean data might be more robust than that obtained using the combined land-ocean data. We used the combined land-ocean record because this quantity in fact yields the change in global mean surface temperature that is conventionally employed in the definition of Earth's climate sensitivity. However the concern arises over systematic errors in the land record from station siting, for example. However recent examination has shown little effect from such siting issues [Rohde et al., 2013]. A more intrinsic question might be whether land surface temperature inherently exhibits a greater response to forcing than ocean temperature, as indicated, for example, by Fasullo [2010]. Nonetheless, as the land temperature contributes only 30% to the global mean temperature we feel confident in our use of the global surface temperature record in the present analysis, although we would not preclude the use of only the ocean surface temperature record in future work.

We might finally observe that equilibrium climate sensitivity should not be viewed as a general property of Earth's climate system but rather as a property of the present climate system exposed only to minor perturbations about an initial steady state. Climate sensitivity specifies only the response of global mean surface temperature to the radiative perturbation, it presents thus only a one-dimensional view of a very rich, multi-dimensional response of the climate system to such a perturbation. Nonetheless, at present even this very limited quantity is highly uncertain, at least a factor of 2 in the 2007 IPCC Assessment. Moreover, climate model studies have shown that climate sensitivity is highly sensitive to parameterizations of sub-grid processes within the limits of present understanding [e.g., Sanderson et al., 2008; Collins et al., 2011]. Consequently any information that can be gained on climate sensitivity from empirical assessments such as the present one must be considered as useful in furthering understanding of the climate system and constraining estimates of this quantity by any approach.

Summary and Conclusions

Principal approaches to determining Earth's climate sensitivity are studies with climate models and empirical determination from temperature change and forcing, either over the historical record or from paleo records. In principle if the models are physically correct, the climate model approach is by far the most comprehensive method, and consequently this approach has been the focus of much investigation, as summarized and assessed in the several IPCC reports and elsewhere. However current climate models rest heavily on assumptions and parameterizations, especially in their treatment of clouds, that are manifested by large differences in the feedbacks and resultant climate sensitivity (Bony et al., 2006; Webb et al., 2006; Soden and Held, 2006; Stevens and Boucher, 2012). For that reason we argue that empirical assessments are of considerable value, and it is in that spirit that we have conducted the present investigation.

Examination of the record of global temperature and forcing by greenhouse gases shows that these quantities have broadly been running in parallel for the major part of over the 20th century, with an average ratio of ca 0.3 K $(Wm^{-2})^{-1}$, Fig.2. Interpretation of this ratio as an integrated transient climate sensitivity is intriguing. However such a low value is generally interpreted as due mainly to the effect of anthropogenic tropospheric aerosols reducing the forcing of greenhouse gases. Accepting this interpretation implies *de facto* that human society has inadvertently been engineering the climate during the whole period. As these aerosols are short-lived in the atmosphere, this interpretation would imply also that future reduction in the emissions of aerosol precursor gases in conjunction with future reductions in CO_2 emissions would give rise to a rapid increase in global temperature as the aerosol offset is reduced.

In this study we examined data for the time period 1970-2010 for which measurements of ocean heat content and global temperature permit calculation of transient and equilibrium sensitivity, provided forcing is known or assumed. For forcing we used the forcing due only to incremental greenhouse gases over this period. Based on satellite observations and records of emissions we argued that the change in aerosol forcing over this period was small, and if anything negative (net cooling influence). Consequently our use only of incremental greenhouse gas forcing in calculating transient and equilibrium sensitivities yields a lower bound to these quantities. Our best estimate lower bounds to these quantities are 0.39 ± 0.07 and 0.54 ± 0.14 K (W m^{-2})$^{-1}$, respectively, equivalent to 1.46 ± 0.26 and 2.02 ± 0.53 K (3.7 W m^{-2})$^{-1}$, where the latter unit permits comparison to commonly presented estimates and assessments of transient climate response and equilibrium CO_2 doubling temperature; the uncertainties represent 1-σ estimates evaluated from uncertainties in forcing, temperature change, and rate of change of ocean heat content. The best estimate for transient sensitivity that we found is at the low end of the range of transient climate response at the time of CO_2 doubling in recent 1% per year climate model experiments, which varies between 1.2 K and 2.6 K temperature increase, with a mean value of 1.9 K. Likewise, our best estimate of the lower bound climate sensitivity essentially coincides with the low end of the "likely" range (central 68% of the probability distribution function) of equilibrium sensitivity given in the 2007 IPCC Assessment.

We also presented quantities that we denoted as lower bounds to the two climate sensitivities, which we calculated as the best estimate minus 1.64 σ, to extend the uncertainty range to encompass all but the 5% tail of the distribution, for the probability distribution function for these quantities taken as normally distributed. For these quantities we obtained for transient and equilibrium sensitivities 0.28 and 0.31 K (W m^{-2})$^{-1}$, respectively, equivalent to 1.03 and 1.16 K (3.7 W m^{-2})$^{-1}$. The lower bound to equilibrium sensitivity calculated in this way exceeds the no-feedback Planck sensitivity, establishing observationally, within the assumptions of this analysis, that feedback in the climate system can confidently be taken as positive.

With respect to an observationally based best central or upper limit estimate of climate sensitivity, we, as others have been as well, are limited by lack of confident knowledge of forcing, specifically the incremental aerosol forcing over the period examined here 1970-2010.

We note however that improvements in monitoring aerosol amount and radiative influence by satellite give hope for the ability to quantify aerosol forcing in the not too distant future, with the resultant ability to yield a best estimate for climate sensitivity, not just a lower bound. This would amount to a major advance in confident understanding Earth's climate system and its susceptibility to perturbations, given the difficulty in determining Earth's climate sensitivity from model calculations, as long recognized (Hansen et al., 1984; Schlesinger, 1988) and more recently underscored by Roe and Baker (2007). In this regard we noted that if the incremental negative aerosol forcing between 1970 and 2010 were as great (in magnitude) as 0.5 W m^{-2}, the transient sensitivity would be S_{tr} = 0.60 K (W $m^{-2})^{-1}$, and the equilibrium sensitivity would be 1.07 K (W $m^{-2})^{-1}$, equivalent to 4.0 K (3.7 W $m^{-2})^{-1}$. As such a high incremental aerosol forcing is unsupported by satellite observations, we consider it therefore highly unlikely that equilibrium climate sensitivity is greater than about 4 K (3.7 W $m^{-2})^{-1}$. As this value is well within the range of current estimates, this result is more important in constraining the upper bound of climate sensitivity than in providing an improved best estimate of this sensitivity.

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Figure legends

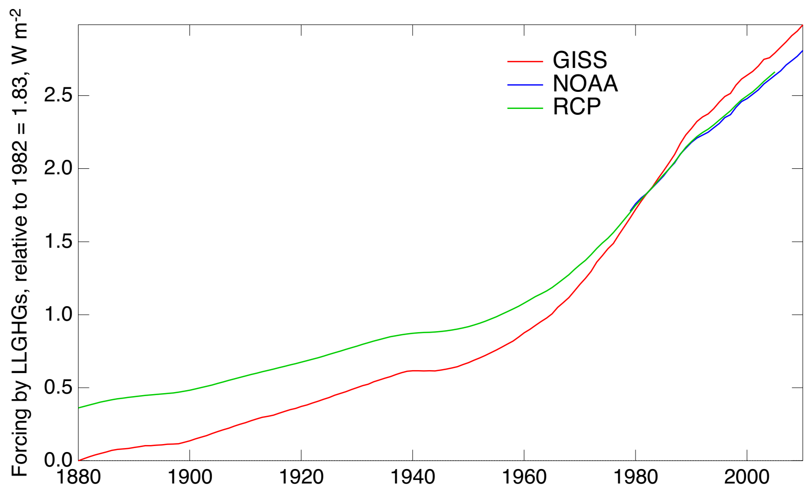
Figure 1. GHG forcing as presented by the Goddard Institute for Space Studies (GISS; http://data.giss.nasa.gov/modelforce/RadF.txt), National Oceanic and Atmospheric Administration (NOAA; http://www.esrl.noaa.gov/gmd/aggi/AGGI_Table.csv) and the Representative Concentration Pathways group (RCP; http://www.pik-potsdam.de/~mmalte/rcps/data/20THCENTURY_MIDYEAR_RADFORCING.xls). All forcings are set equal at 1982 to permit comparison.

Figure 2. Correlation of global temperature and GHG forcing. Temperature anomaly data are HadCrut3 (Brohan et al., 2006, as extended at http://www.cru.uea.ac.uk/cru/data/temperature/). Forcing (blend of RCP and NOAA as discussed in text) is relative to preindustrial. Ratio of scales of two vertical axes was set by slope of graph of ΔT_S vs. forcing.

Figure 3. Forcing by LLGHGs and non LLGHG forcing over the time period 1970-2010 as given by the GISS and blended RCP-NOAA data sets.

Figure 4. Graph of temperature anomaly vs. forcing by LLGHGs for the years 1970-2010 (indicated by color). Forcing is average of GISS and blended RCP-NOAA, relative to 1970, Figure 3. Slope $S_{tr} = 0.39 \pm 0.03$ K/(W m^{-2}), where the 1-σ uncertainty is based only on the uncertainty in the fit; forcing is relative to 1970; temperature anomaly HadCrut3 is relative to base period (1961-1990). Correlation coefficient $r^2 = 0.80$.

Figure 5. Heat content of the world ocean to depth of 2000 m. Slope ($0.48 \pm 0.02 \times 10^{22}$ J yr^{-1}) of linear fit (blue) to data for years 1970 - 2008, indicated by arrows, corresponds to heating rate relative to the area of the planet $N = 0.30 \pm 0.01$ W m^{-2}. Data from Levitus et al. (2012).



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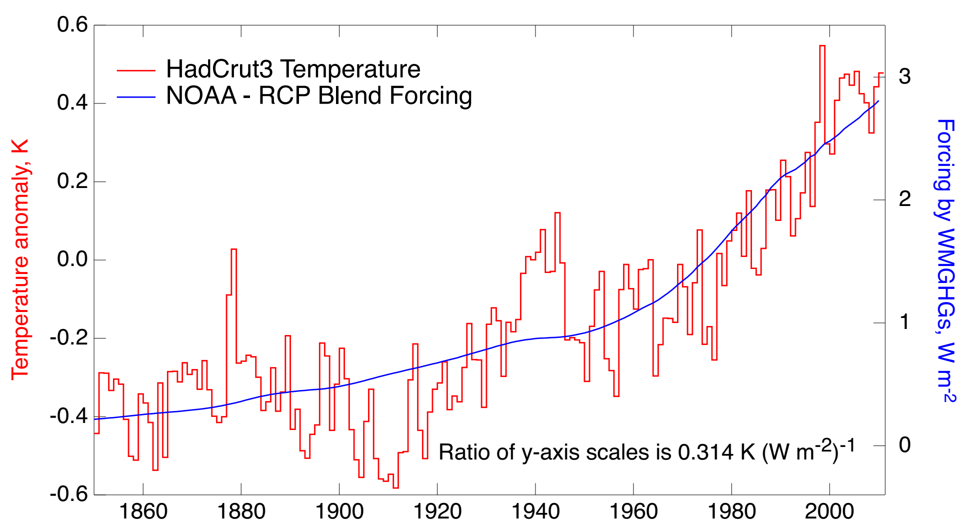

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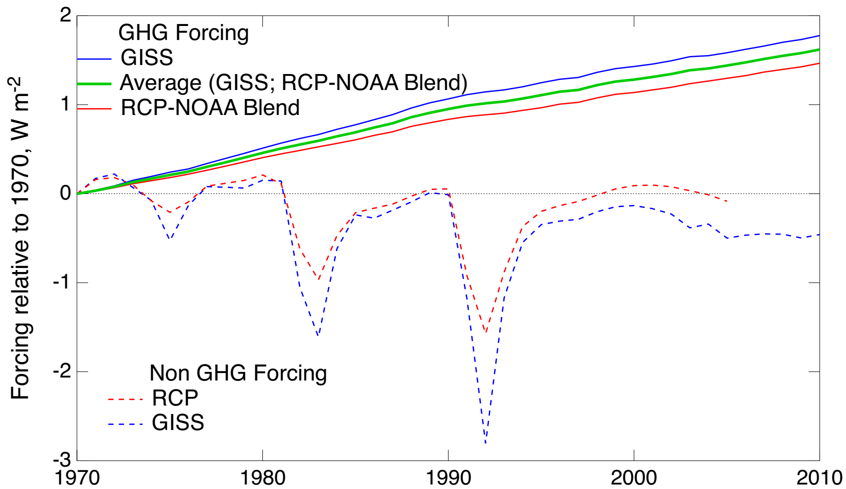

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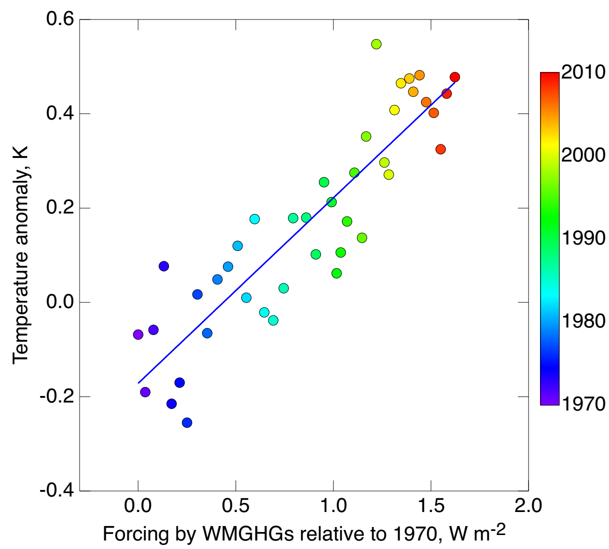

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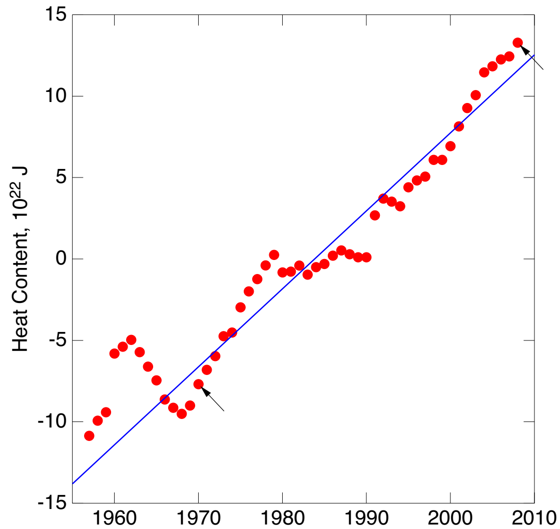

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