RESOURCE LETTER (Submitted to American Journal of Physics, 2018)

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ABSTRACT. Human activities over the past 200 years have increased Earth’s greenhouse effect by about 1% relative to the radiative fluxes that drive the climate system. This Resource Letter introduces the physics of this intensified greenhouse effect and of the processes that govern resultant change in climate and provides resources for further study. Context for this examination is provided in a companion Resource Letter (GECC-1) that examined the radiative fluxes that comprise Earth’s climate system and the present-day greenhouse effect. The increase in global temperature and other changes in climate resulting from the intensified greenhouse effect are of great societal concern. Developing prognostic capability to determine these responses to the small perturbations that constitute the intensified greenhouse effect to an accuracy that would be useful to inform policymaking is the major challenge facing climate scientists today.

I. INTRODUCTION

As described in the companion Resource Letter (GECC-1), the so-called “greenhouse effect” results in an increase in global mean surface temperature (GMST) that is about 32 K greater than what it would otherwise be for the same planetary absorption of solar radiation and is thus a
major feature of Earth’s present climate. The examination in GECC-1 of the magnitudes of the several energy fluxes that characterize Earth’s climate system, the physical processes that give rise to the greenhouse effect, and the magnitude of the greenhouse effect provides context and sets the scene for the examination here of the increases of atmospheric amounts of greenhouse gases (GHGs) and other changes in atmospheric composition due to human activities over the Anthropocene epoch,\textsuperscript{1-3} the period of time over Earth’s biogeochemistry and climate have been appreciably affected by human activities, roughly the past 200 years, and the consequences of prior and prospective changes in climate. The emphasis here will be on change in GMST. It is commonly assumed, and generally found in studies with large-scale climate models, that changes in other climate properties, such as precipitation, will be linear in the change in GMST, but with recognition that the changes in climate variables will not be spatially uniform.

Sources and References. As indicated in Resource Letter GECC-1, a broad and detailed examination of present understanding of the climate system and climate change is given in the several Assessment Reports of Working Group I of the Intergovernmental Panel on Climate Change (IPCC), especially the two most recent reports.\textsuperscript{4,5} These reports, which are freely downloadable on the web, are essential references for any student of climate and climate change. Resource Letter GECC-1 called attention also to several texts that might serve as an introduction to the climate and to the physics of climate. A hypertext history of the development of understanding of the greenhouse effect and climate change more generally by Weart,\textsuperscript{6} which supplements his very readable book, may be accessed at http://www.aip.org/history/climate/index.htm. The monograph by Brönnimann\textsuperscript{7} deals with many aspects of natural and anthropogenic climate change over the past 500 years. Citations to the primary and secondary literature given in this Resource Letter are intended to be of historical and
pedagogical value, as well as to lead the reader to the research literature without attempting to be exhaustive.

Geophysical data. Data sets particularly pertinent to climate change over the Anthropocene in addition to those given in Resource Letter GECC-1 are time series for components of global mean radiative forcing and for components of the global atmospheric CO$_2$ budget; see also citations to data sources given in captions to Figure 1 and Table 2.

Organization of this Resource Letter. Section II quantifies the increases in amounts of GHGs over the Anthropocene, introduces the radiative forcing concept, and quantifies forcings by incremental GHGs and other forcing agents over the Anthropocene. Section III focuses more explicitly on forcings by incremental GHGs and on the atmospheric residence times of the key GHGs. Section IV examines climate system response to forcings, introduces the climate sensitivity concept, and examines estimates of Earth’s climate sensitivity and present uncertainty in these estimates. Section V examines the expected change in global temperature that would be expected based on estimates of forcings over the Anthropocene together with estimates of Earth’s climate sensitivity, again with associated uncertainties and examines implications for prospective future change in global temperature. A summary is presented in Section VI.

Supplementary Notes (SN) in the online version of this Resource Letter provide detail or lend perspective, as follows.

SN1. Calculation of radiative forcing by an incremental greenhouse gas.
SN2. Can forcings from increases in GHGs be measured?
SN3. The budget and adjustment time of incremental atmospheric CO$_2$.
SN4. Treating global temperature response as linear in the perturbation.


8. Supporting Information to IPCC AR5 Chapter 8, Annex II: Climate System Scenario Tables, Table AII.1.2, “Historical effective radiative forcing (ERF) (W·m⁻²), including land use change (LUC)” Published 31 January 2014; corrected xls-sheets as of 20 August 2014. http://www.climatechange2013.org/images//report/WG1AR5_AIIISM_Datafiles.xlsx.


II. RADIATIVE FORCING OF CLIMATE CHANGE

The anthropogenic influences on the planetary radiation budget and responses of the climate system to perturbations are the key questions that face the climate change research community today. Well established by measurements, atmospheric mixing ratios of several long-lived gases that contribute to Earth’s greenhouse effect have increased substantially over the past 250 years because of human activities [see Figure 1]. Mixing ratios prior to contemporary measurements are from glacial ice cores, mainly in Antarctica, for which the record extends to 800,000 years before the present. The dataset initiated in 1958 by Charles David Keeling¹⁰,¹¹ at Mauna Loa, Hawaii, isolated from local sources and sinks and thus representative of the Northern Hemisphere, reveals the annual cycle of draw down and release of CO₂ by terrestrial vegetation. This dataset has become an icon of the anthropogenic influence on GHG concentrations. The increase in abundance of carbon dioxide and methane over the Anthropocene is comparable to
the increase between the last glacial maximum and the Holocene, the current temperate epoch of Earth’s geological history. Shown on the right-hand axes of the several panels in Figure 1, are the perturbations in the longwave radiation budget of the planet that would result from the increases in the mixing ratios of the several gases. This hypothetical increase in the net rate of uptake of energy by the climate system (or more generally, any such change in Earth’s energy budget that is imposed from outside the climate system), is denoted a *forcing*; like the energy fluxes themselves, forcings are commonly given normalized to the area of the planet, with unit $W \cdot m^{-2}$. The sum of the several forcings, during the Anthropocene because of the increase in abundance of GHGs, leads to expectation of a resultant increase in the greenhouse effect and consequent increase in Earth’s surface temperature. The increases in atmospheric abundances of the so-called long-lived GHGs, most importantly carbon dioxide (CO$_2$), but also including methane (CH$_4$), nitrous oxide (N$_2$O), and chlorofluorocarbons (CCl$_2$F$_2$, CCl$_3$F) are at the core of the present concern over anthropogenic global warming, or more broadly over anthropogenic climate change.
Figure 1. Atmospheric mixing ratios (left axes) and resulting global-mean perturbations in radiative budget (radiative forcings, right axes) of carbon dioxide, methane, and nitrous oxide over (a) the last glacial-interglacial cycle, 150,000 years; (b) the Holocene, 10,000 years; (c) the Anthropocene, 250 years, and (d) contemporary measurements at Mauna Loa, Hawaii. Measurements from Antarctic ice cores in (b) and (c) are shown as symbols with different colors for different studies; contemporaneous atmospheric measurements are shown by red curves. (a) CO$_2$, Ref. [12]; CH$_4$, Ref. [13]; (b) and (c) modified from the Fourth IPCC Assessment Report (Ref. [5] §6.4.1.1); (d) monthly mean data from NOAA (ftp://aftp.cmdl.noaa.gov/data/trace_gases/) and, for CO$_2$, Scripps Institution of Oceanography (http://scrippscio2.ucsd.edu/data/atmospheric_co2/primary_mlo_co2_record), updated from Ref. [14].

The magnitude of forcing by the several long lived GHGs, about 3 W·m$^{-2}$ at present relative to the preindustrial atmosphere, is some two orders of magnitude less than the global and annual mean outgoing longwave flux at the top of the atmosphere (TOA) or than the corresponding average downwelling or upwelling longwave fluxes at the surface (Resource Letter GECC-1, Figure 2). This disparity serves to justify treating this forcing as a perturbation on the climate system. Moreover, the spatial and temporal variations in the several fluxes inherent in the climate system (Resource Letter GECC-1, Figure 3) greatly exceed the magnitude of the perturbation.
due to increased amounts of atmospheric GHGs, challenging the capability to measure the radiative effects of the incremental GHGs (SN2). Consequently, these radiative effects are not determined by measurement but are calculated, based on the absorption spectra of the several gases and the atmospheric state.

Over the Anthropocene, industrial activities have also resulted in increases in the amounts of aerosols in the troposphere. Aerosol particles scatter shortwave radiation and increase the reflectivity and persistence of clouds; increases in aerosol loadings would be expected to reduce the absorption of shortwave radiation incident on the planet, thereby changing Earth’s radiation budget (i.e., exerting a forcing) in a direction opposite to the increase in net energy due to increased GHGs. The magnitudes of the radiative perturbations due to anthropogenic aerosols are comparable to those of the GHGs, but the perturbations are much less well understood and quantified; absorption of shortwave radiation by soot particles lends further complexity to the aerosol forcing. Incremental tropospheric aerosols are produced largely due to fossil fuel combustion through direct emission and atmospheric reactions involving sulfur and nitrogen oxides emitted as combustion byproducts. A major difference between the influences of incremental GHGs and incremental aerosols arises from the vast difference in atmospheric residence times of the GHGs (decades to centuries) and the tropospheric aerosols (about 10 days, through removal mainly by precipitation), the implications of which are examined in §V. Quantitative understanding of the climate system response to perturbations over the Anthropocene and to responses to future changes in atmospheric composition must take all of these perturbations into account. As the aerosol influences are not well understood, they are the subject of much current research. Present best estimates by the IPCC of global-mean forcings of Earth’s radiation budget over the Anthropocene epoch are shown in Figure 2. The sign
convention is that forcings, which increase the net energy of the planet, are denoted as positive and those which decrease the net energy are negative; thus, forcings by incremental GHGs are positive and forcings by incremental aerosols are in the aggregate negative. As discussed in §IV and SN4, change in GMST in response to forcings can be considered in good approximation to be linear, at least to first order; that is for a forcing of a given type the response is proportional to the strength of the forcing, and forcings of different types are algebraically additive. Thus, the negative forcing by anthropogenic aerosols must be considered as offsetting some fraction of the positive forcing by GHGs.

In addition to the gradual, longer-term perturbations due to the build up of atmospheric GHGs and aerosols over the Anthropocene, the total forcing has been punctuated by intermittent large negative forcings arising from major volcanic eruptions. These eruptions inject sulfur dioxide into the stratosphere; sulfuric acid aerosol formed from this sulfur dioxide scatters incident shortwave radiation, thus increasing the planetary albedo. The forcing from a given volcanic eruption decreases gradually over a several-year period as the aerosol is removed from the stratosphere as stratospheric air is exchanged with that of the troposphere; once in the troposphere the aerosol is rapidly (weeks) removed from the atmosphere, mainly by precipitation.
Figure 2. Radiative forcings, relative to 1750, by key climate influencing agents over the Anthropocene epoch, left, and for 2011, numerical values and bars at right. Bar denoting forcing by tropospheric aerosols represents the sum of direct aerosol interaction with radiation (Aer-Rad Int.) and indirect interaction through aerosol cloud interactions (Aer-Cld Int.). Uncertainties on bars represent central 90% range of likelihood distribution. Large negative spikes denote forcing by aerosols, mainly in the stratosphere, resulting from major volcanic eruptions; magnitudes of early volcanos are truncated. Modified from IPCC, Figure 8.18 by addition of numerical values from Table 8.6 and of names of major volcanos.

Shown at the right of Figure 2 are the best estimates of the values of present (2011) forcings by the several forcing agents (bars) and of the associated uncertainties (I-beams denoting central 90% of likelihood distribution) given by the 2013 IPCC assessment. The relatively small uncertainties associated with forcings by the incremental GHGs contrast to the quite large uncertainties associated with the incremental tropospheric aerosols reflected in the large uncertainty range given in the Assessment [-0.09 to -1.88 W·m⁻²]. Because of the additivity of forcings, the best estimate of total forcing carries with it the uncertainty associated with the aerosol forcing. If the aerosol forcing is at the low-magnitude end of the indicated range, then the aerosol is offsetting only a small fraction of the forcing by GHGs, and the total forcing is at the high end of the indicated range, 3.33 W·m⁻². If, however, the aerosol forcing is at the high-
magnitude end of the uncertainty range, the offset of GHG forcing by aerosol forcing is quite substantial, with the total forcing 1.13 W·m⁻². The resultant uncertainty in total forcing is thus a factor of 3. As discussed in §IV this uncertainty in total forcing over the Anthropocene epoch, has major implications with respect to interpretation of climate change over this period.


III. THE INTENSIFIED GREENHOUSE EFFECT

The foregoing overview of the climate system, with emphasis on the processes that control Earth’s radiation budget, permits examination of the anthropogenic perturbations to this budget. The important drivers of this change are increases in atmospheric infrared-active gases over the Anthropocene, shown in Figure 1, and increases in tropospheric aerosols. This section examines the changes in amounts of greenhouse gases and the resultant forcings. Determination of the changes in Earth’s radiation budget due to perturbations in composition requires knowledge of the increments in atmospheric abundances of the pertinent substances and evaluation of the resulting perturbations in radiation. The forcings by incremental greenhouse gases are, to a good approximation, linear in the amount of incremental gas. It is thus possible to separate the calculation of the forcing into two components: 1) determination of the anthropogenic perturbation to the amount of greenhouse gas, and 2) determination of the change in TOA flux per increment of the gas. This separation simplifies the problem and permits attribution of the sources of uncertainty to one or the other component.

A. Anthropogenic increments in greenhouse gases

As shown in Figure 1, the amounts of carbon dioxide, methane, and nitrous oxide, which had been relatively constant for several thousand years, have increased substantially and systematically over the past two centuries, a consequence of human activities. These increments
are firmly established by measurements in ice cores and, for more recent times, direct atmospheric measurements. The increase in the amount of CO₂ in the first half of the 19th century was due mainly to deforestation; subsequently this increase has been increasingly dominated by emission from fossil fuel combustion, which now accounts for almost 90% of total CO₂ emissions. The increases in methane and nitrous oxide are due to a mix mainly of agricultural activities and energy production, Table 1. Additionally, chlorofluorocarbons CCl₂F₂ and CCl₃F, which are also strongly infrared-active, have been introduced into the atmosphere through their use as refrigerants and propellants. Although production of these gases has been greatly curtailed because of their role in depletion of stratospheric ozone, these gases remain present in the atmosphere. These changes in atmospheric composition have resulted in a change Earth’s radiation budget, mainly in the longwave, and thus a radiative forcing of climate change.

<table>
<thead>
<tr>
<th>Gas</th>
<th>CO₂</th>
<th>CH₄</th>
<th>N₂O</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fossil-fuel related &amp; cement production</td>
<td>87</td>
<td>29</td>
<td>3</td>
</tr>
<tr>
<td>Net deforestation &amp; Biomass burn</td>
<td>13</td>
<td>11</td>
<td>12</td>
</tr>
<tr>
<td>Total Agriculture &amp; Fertilizer</td>
<td>38</td>
<td>81</td>
<td></td>
</tr>
<tr>
<td>Animal husbandry</td>
<td>27</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rice cultivation</td>
<td>11</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Landfills</td>
<td>23</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Human excreta</td>
<td>3</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 1. Fractional contributions by source category, per cent, to total current anthropogenic emissions of CO₂, CH₄, and N₂O. Based on emissions estimates provided by IPCC, Chapter 6.
B. Radiative forcing by incremental greenhouse gases

The radiative forcing due to an increase in the amount of a greenhouse gas in the atmosphere is the net decrease in outgoing longwave radiation at the TOA that would result from such an increase prior to any increase in Earth’s surface temperature that would counteract the radiative imbalance imposed by this increase. This forcing is determined by radiation transfer calculations that rely on the spectroscopic properties of the several gases. Such radiation transfer calculations are carried out with global-scale models that have realistic distributions of temperature, clouds, and water vapor. Integration over space and over an annual cycle (or perhaps several years to average out fluctuations) yields the forcing for a given increment of greenhouse gas or gases. The example calculation of radiative forcing in SN1 explicitly illustrates the dependence of this forcing on the vertical temperature structure of the atmosphere.

Calculations such as those in SN1 but carried out globally and over annual and diurnal cycles, have led to development of simplified expressions for the global mean forcings of the several long-lived GHGs given in Table 2. The dependence of CO$_2$ forcing on mixing ratio shown in the table as logarithmic is sublinear in mixing ratio (slope decreases with increasing value of argument), as the increase in absorption is due only to increased absorption in the wings of the lines, the line centers being saturated. For CH$_4$ and N$_2$O the lines are not quite so saturated, resulting in an increase in forcing with increasing mixing ratio that is less sublinear than that of CO$_2$ and which is commonly represented as a square-root dependence. For the chlorofluorocarbons, CCl$_2$F$_2$ and CCl$_3$F, which are entirely anthropogenic and for which the base-case mixing ratio is zero, the forcing is linear in mixing ratio and is much stronger per incremental mixing ratio than for CO$_2$. Although commonly given expressions for forcing by the incremental GHGs are sublinear, the forcings to date, relative to the preindustrial climate, are
approximately linear, Figure 3, although sublinear expressions would be needed for wider ranges of mixing ratios. The linearized expressions given in Table 2 explicitly show the much stronger forcing per increment of mixing ratio for chlorofluorocarbons, followed by N₂O and CH₄ than for CO₂. It is this variation in forcing per incremental amount of gas in atmosphere, which gets entirely obscured by the several different functional forms, that is much more the take-home message than any nonlinear dependence on mixing ratio.

![Figure 3](image.png)

**Figure 3.** Global average radiative forcing by long-lived greenhouse gases as a function of their mixing ratios (parts per million, ppm; billion, ppb; trillion, ppt) shown on abscissa beginning at preindustrial values and extending approximately to present values. Red, forcing calculated by approximate formulas. Blue lines denote linear dependence of forcing on mixing ratio. Note differences in horizontal and vertical scales for the several gases.

<table>
<thead>
<tr>
<th>Trace Gas</th>
<th>Preindustrial mixing ratio, x₀, ppb</th>
<th>2016 mixing ratio, x, ppb</th>
<th>Expression for forcing, W·m⁻²</th>
<th>2016 forcing, W·m⁻²</th>
<th>Slope of linear fit, W·m⁻²·ppb⁻¹</th>
<th>Adjustment time, yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>CO₂</td>
<td>278 × 10³</td>
<td>403 × 10³</td>
<td>5.35(ln x – ln x₀)</td>
<td>1.99</td>
<td>0.0000159</td>
<td>40 – 50</td>
</tr>
<tr>
<td>CH₄</td>
<td>700</td>
<td>1843</td>
<td>0.036(√x – √x₀)</td>
<td>0.51</td>
<td>0.00051</td>
<td>12</td>
</tr>
<tr>
<td>N₂O</td>
<td>270</td>
<td>329</td>
<td>0.12(√x – √x₀)</td>
<td>0.19</td>
<td>0.0035</td>
<td>120</td>
</tr>
<tr>
<td>CCl₂F₂</td>
<td>0</td>
<td>512</td>
<td>0.33(x – x₀)</td>
<td>0.16</td>
<td>0.33</td>
<td>100</td>
</tr>
<tr>
<td>CCl₃F</td>
<td>0</td>
<td>230</td>
<td>0.25(x – x₀)</td>
<td>0.06</td>
<td>0.25</td>
<td>45</td>
</tr>
</tbody>
</table>

**Table 2.** Mixing ratios, forcings, and adjustment times of long-lived greenhouse gases. Mixing ratio x is in parts per billion, ppb; subscript 0 denotes preindustrial. Expressions for forcings from Ref. [18]. Linear approximations are shown in Figure 3. Mixing ratios and forcings are from NOAA annual greenhouse gas index [http://www.esrl.noaa.gov/gmd/aggi/aggi.html](http://www.esrl.noaa.gov/gmd/aggi/aggi.html); accessed 2017-11-17. Adjustment times from IPCC (Ref. [4]; §8.3) except for CO₂ (Refs. 20, 21; SN3).
Because of the importance of forcing by GHGs as a driver of climate change, and because forcing cannot be directly measured (SN2) but can be determined only by radiative transfer calculations, such calculations assume considerable significance. A measure of the accuracy of such calculations comes from comparisons of forcings as calculated by multiple models. The global mean forcing that would result from doubling the atmospheric mixing ratio of CO₂, as calculated for a cloud-free atmosphere by several radiation transfer models and by the radiation codes of several general circulation climate models, was found to exhibit quite small spread (standard deviation 0.07 W·m⁻² or 1.6%). This small spread is indicative of the confidence that can be placed in such calculations. However, as the comparison was conducted for a cloud-free planet in order to avoid complications arising from differing treatments of clouds in climate models, the resulting forcing, about 4.5 W·m⁻², substantially exceeds the 3.7 W·m⁻² for forcing by doubled CO₂ obtained with the widely used expression in Table 2. Moreover, it has become appreciated in the past several years that the introduction of a greenhouse gas into the atmosphere would likely induce changes in atmospheric thermal structure and in turn clouds, thereby modifying the radiative impact of the incremental greenhouse gas from what it would be with static atmospheric structure. As this adjustment occurs rapidly (days to weeks), well prior to appreciable response of global surface temperature to the forcing, the resultant radiative changes are more usefully viewed as altering the forcing rather than as a response of the climate system to the forcing. The resulting forcing has been denoted the “adjusted forcing.” The inter-model spread in adjusted forcing by doubled CO₂ is much greater than that obtained for the pure radiative forcing in the absence of clouds and without accounting for adjustments. For the 23 climate models that participated in the 2013 IPCC model comparisons, the adjusted forcing exhibited a mean of 3.44 W·m⁻², and 90% uncertainty range [± 0.84 W·m⁻²], equivalent to
This spread in estimated forcing is close to the IPCC \(^4\) estimate of 90% uncertainty range in forcing by long-lived GHGs [± 20%]. Of course, the question always remains whether spread in model calculations is an accurate estimate of uncertainty, but it is in some sense a lower bound to the uncertainty that can be ascribed to present estimates.

One further greenhouse gas that should be mentioned is tropospheric ozone, which is produced by atmospheric chemical reactions involving nitrogen oxides emitted as byproducts of fossil fuel combustion and natural and anthropogenic hydrocarbons. Concentrations of ozone have increased substantially over the past 250 years. As production is localized to areas of high emission density and exhibits a rather strong seasonal cycle, and as excess atmospheric ozone is rather short lived (months), the gas is rather nonuniformly distributed spatially and temporally. The best estimate of the greenhouse forcing by incremental tropospheric ozone is about 0.4 ± 0.2 \(W \cdot m^{-2}\); a forcing of this magnitude, although not negligible as an agent of climate change, is small relative to forcing by the incremental long-lived GHGs. The large relative uncertainty, [± 50%] (central 90% of the likelihood distribution), is due much more to uncertainty in the amount and distribution of the increase in tropospheric ozone over the Anthropocene than to uncertainties associated with radiative transfer.

Two additional points should be noted regarding forcings. First, as indicated in §II and examined further in §IV, the forcing by increases in GHGs over the Anthropocene represents the amount by which the longwave energy leaving the planet would have decreased, relative to the preindustrial steady state, in the absence of any increase in global temperature and resultant increase in longwave emission (and/or decrease in shortwave absorption) that would offset the imposed imbalance. As discussed in SN1 of Resource Letter GECC-1, a forcing of 3 \(W \cdot m^{-2}\), if not compensated by restoration of the planetary energy balance resulting from increase in surface
temperature, would result in a planetary warming rate much greater than has been experienced over the Anthropocene. Thus, the change in net irradiance at the top of the atmosphere (TOA) in response to this applied forcing (i.e., the planetary energy imbalance) must be well less than the applied forcing, and this would apply in general for the planetary response to any applied forcing.

A second important point about the forcing by increases in GHGs over the Anthropocene is that this forcing represents a perturbation of about 1% of the total global and annual mean outgoing longwave radiation at the TOA. The requirement is thus to determine the consequences of a 1% change in a quantity that is highly variable spatially and temporally to some desired accuracy, say 25%. This is a key reason why quantifying the effects of the increases in GHGs on Earth’s climate has been and continues to be such a challenge.

In summary, current estimates of radiative forcing by incremental GHGs, at present and as a function of time over the Anthropocene, rest on radiative transfer calculations, averaged globally and over the annual cycle. They take into account the optical properties of the gases and other controlling influences, importantly temperature and its vertical structure, spectral interference by other GHGs including water vapor, and interference by clouds (e.g., Ref. [25]). Although based on well-understood physics, such calculations are not trivial, and likewise calculation of global average forcing is not trivial. Temperature exhibits seasonal and more rapid variation, affecting greenhouse gas forcing directly and indirectly through radiative interactions with water vapor and clouds. Consequently, even though the incremental amount of the gas is fairly uniformly distributed in the global atmosphere on account of the long residence times of these gases, forcings by incremental GHGs are, in general, dependent on location, season, time of day, cloudiness, and the like. All these situational dependences must be accounted for in calculating
the global averages of the several forcings. Still more complicated is modeling of forcings by
tropospheric aerosols, whose properties and distributions in the atmosphere are highly
nonuniform and whose interactions with clouds are not yet well understood. It is estimates of
forcings obtained in this way, together with uncertainties inferred from inter-model spread and
observational constraints, that were examined in the 2013 IPCC Assessment and which are
summarized in Figure 2.

C. Adjustment times of forcing agents

A further important property of forcing agents (greenhouse gases, aerosols) influencing their
impacts on climate change is their residence times in the atmosphere. These residence times are
governed by chemical, physical, and, for CO₂, biological processes that remove the substances
from the atmosphere. Removal of greenhouse gases methane, nitrous oxide, and
chlorofluorocarbons takes place largely by gas-phase chemical reactions on time scales of a
decade or so (CH₄) to a century or so (N₂O, CFCs), Table 2. Aerosol particles are removed from
the atmosphere mainly by precipitation and, especially for larger particles (diameter ≥ 3 μm)
gravitationally and/or and inertially induced deposition to the surface, on a much shorter time
scale, about a week. The situation with anthropogenic CO₂ is more complicated, because of the
abundant natural background of CO₂ with sources and sinks that greatly exceed the source
strength of anthropogenic CO₂. It is thus necessary to distinguish the mean atmospheric
residence time of CO₂, from what is termed the adjustment time of incremental CO₂, the time
constant that would characterize the rate of removal of incremental CO₂ in the atmosphere
relative to the preindustrial amount (e.g., Ref. [4], Glossary, p. 1457). This adjustment time is the
quantity that characterizes the persistence of the intensified greenhouse effect of anthropogenic
emissions of CO₂. The role of natural sources and sinks of CO₂ is manifested in the annual
fluctuations of atmospheric CO₂ mixing ratio seen, for example, in the Mauna Loa record, Figure 1. From estimates of the rate of uptake of CO₂ into the terrestrial biosphere and ocean (Ref. [5], Figure 7.3) the mean residence time of atmospheric CO₂ (amount in the atmosphere divided by removal rate) is estimated at about 3 years. However, the residence time reckoned in this way is inappropriate for consideration of the adjustment time of incremental CO₂. From its definition, it is seen that the adjustment time of incremental CO₂ is a hypothetical quantity, as it cannot be directly measured. However, it can be inferred by modeling of the planetary carbon cycle²⁶,²⁷ or from considerations of the budget of excess atmospheric CO₂, (Refs. 20, 21; SN3). Budget considerations indicate an adjustment time for CO₂ at present of about 45 years, increasing somewhat from 1958 to the present and thus likely to continue to increase in the future.


IV. CLIMATE SYSTEM RESPONSE TO FORCING; CLIMATE SENSITIVITY

The key question facing the climate change research community is how much global temperature would be expected to increase for a given increase in CO₂ or other GHGs in the atmosphere. Closely related is attribution of the observed increase in global temperature to the increase in GHGs and whether the magnitude of this increase is consistent with expectation. Yet another key question is the time scale over which such temperature response would occur. Change in global mean surface temperature (GMST) is commonly taken as the principal indicator of climate change under the assumption that changes in other components of the climate system would scale linearly with change in GMST. This hypothesis is borne out for global-annual quantities, such as precipitation amount, in studies with global climate models. Computer representations of the key processes of the climate system permit examination of the consequences of perturbations in amounts of greenhouse gases and aerosols or other perturbations (realistic or intentionally unrealistic) such as changing the solar constant. Locally, and over shorter time periods, changes can be quite nonlinear, especially those associated with the ice-liquid water transition. Here the focus is restricted to GMST as an indicator of climate change on a global scale.

Recognition that a forcing would be expected to induce a change in global temperature leads to definition of a quantity commonly denoted as Earth’s equilibrium climate sensitivity, the amount by which GMST would change, for long times, in response to a sustained global mean forcing $F$, normalized to that forcing,

$$S_{eq} = \frac{\Delta T_s}{F} \text{ or equivalently } \Delta T_s = S_{eq}F$$ (1)

Equation (1) may also be viewed an explicit statement of the linearity of response of GMST to forcing, at least to first order in the perturbation; see SN4. Determination of this so-called
equilibrium sensitivity is a longstanding goal and challenge to the climate research community. It is important here to qualify use of the word “equilibrium” in the context of climate sensitivity. It should be emphasized Earth’s climate is by no means an equilibrium system. A state of dynamic equilibrium is characterized by detailed balance (equal and opposite fluxes) on all paths. This condition is decidedly not met by Earth’s climate system, which is characterized by shortwave radiation in and longwave radiation out. Nonetheless the term “equilibrium climate sensitivity” is in widespread use; hence the subscript “eq.”

Importantly, the equilibrium sensitivity is pertinent to the changes in surface temperature that characterize the response of Earth’s climate system to perturbations generally, not just to the hypothetical situation of a new steady state. Let the $N$ be the net energy flux into the system. For the unperturbed system, this net energy flux consists of two terms, the absorbed shortwave flux $Q_0$ and the emitted longwave flux $E_0$, where the subscripts 0 denote the initial, unperturbed state. In this initial state

$$N_0 = Q_0 + E_0 = 0,$$  \hspace{1cm} (2)

cf. Eq. (1) of Resource Letter GECC-1. The response of the climate system to a perturbation is illustrated schematically in Figure 4 for the example of a hypothetical step-function perturbation, the application of a constant forcing $F_{\text{app}}$ commencing at time $t_0$, shown in the bottom panel. Here “external” means that the perturbation is not a part of the climate system, for example the addition of an incremental amount of a greenhouse gas into the atmosphere or, hypothetically, an increase in the solar constant. The change in radiation budget due to the forcing is to be distinguished from the change in the budget that would result from climate system response to the perturbation. With the sign convention that positive forcing $F$ denotes an increase in the net energy flux into the system, then initially, at time immediately after $t_0$, the rate of change of the
heat content of the planet, \( N \), previously 0, is abruptly increased by the forcing (middle panel of Figure 4),

\[
N(t_0+) = Q_0 + E_0 + F_{\text{app}} = F_{\text{app}}. \tag{3}
\]

The positive net energy flux results in a gradual increase in the heat content of the planet and in turn the surface temperature, top panel. As the temperature increases, the climate system responds by increasing the emitted longwave radiation (decreasing \( E \); recall the sign convention that \( E \) is the negative of the emitted longwave flux) and/or by increasing reflected shortwave irradiance from the planet (decreasing \( Q \), the absorbed shortwave flux). Both responses would, over time, decrease the net downward radiative flux at the TOA, \( N \), offsetting the imposed forcing. This response of the climate system to an imposed forcing is the reason that after time, the energy imbalance of the planet is much less than the applied forcing. A simple, but considerably more realistic, representation of the time response of the climate system to perturbations is given by a two-compartment model of the climate system.\(^{28-32}\) This model exhibits two time constants, a short time constant of about 8 years characteristic of the response of the ocean mixed layer and a long time constant of about 500 years, characteristic of the response of the deep ocean.

Figure 4. Conceptual depiction as a function of time of response of climate system to step-function forcing at time \( t_0 \). \( F_{\text{app}} \) is the applied forcing; \( \Delta T_s \) is the change in surface temperature; \( N \) is the rate of change of heat content with time or, equivalently, the net radiative flux imbalance at the top of the atmosphere, TOA.
Under the assumption (SN4) that the decrease in net downward flux at the TOA is linear in the change in surface temperature, then over time subsequent to imposition of the perturbation,

\[ N(t > 0) = F_{\text{app}} + \left( \frac{\partial Q}{\partial T_{s0}} + \frac{\partial E}{\partial T_{s0}} \right) \Delta T_s(t), \]

where the subscript 0 denotes that the derivatives are evaluated at the initial state, consistent with treatment of effect of the perturbation to first order. At long time such that a new steady state is reached, the net flux into the system \( N(t \rightarrow \infty) \rightarrow 0 \). At such time

\[ \Delta T_s(\infty) = -\left( \frac{\partial Q}{\partial T_{s0}} + \frac{\partial E}{\partial T_{s0}} \right)^{-1} F_{\text{app}} \]

so that the identification can be made, cf. Eq. (1):

\[ S_{\text{eq}} = -\left( \frac{\partial Q}{\partial T_{s0}} + \frac{\partial E}{\partial T_{s0}} \right)^{-1}. \]

Eq. (6), which formally relates the equilibrium climate sensitivity to the derivatives of absorbed shortwave flux and emitted longwave flux with respect to surface temperature, is useful in defining feedbacks in the climate system and as a tool for examining contributions to climate sensitivity in three-dimensional global climate models. The feedback terminology, which derives from its use in analyzing electronic circuits, was introduced into the climate change literature by Hansen et al.\textsuperscript{33} An important insight from Figure 4, and the result holds generally, is that because of climate system response to forcing, a forcing is not a quantity that can be determined by measurement, for example by measurement of the energy imbalance at the TOA. Rather it is a quantity that can be obtained only by calculation based on knowledge of the perturbation in atmospheric composition, taking into account the state of the atmosphere.
The identification in Eq (6) together with Eq (4) shows that the equilibrium sensitivity is a useful measure of climate change not just when the system has attained a new steady state, but generally (e.g., Forster\textsuperscript{34}). Solving Eq (4) for $\Delta T_s(t)$ yields

$$\Delta T_s(t) = S_{eq}[F(t) - N(t)] \quad (7a)$$

where $\Delta T_s(t)$ and $F(t)$ are defined relative to an initial steady state. Still more generally, for two states of the climate system, neither of which is at steady state,

$$\Delta T_s = S_{eq}(\Delta F - \Delta N), \quad (7b)$$

where all $\Delta$’s are calculated between the two climate states. These expressions, which rest on conservation of energy in the climate system and which therefore might be considered “the simplest imaginable parameterization of climate system response to radiative forcing,”\textsuperscript{35} are central to interpretation of climate change. In contrast to forcing, the energy imbalance of the climate system $N$ is a quantity that can be measured, in principle at the TOA from satellites and in practice as the rate of change of global heat content, the great majority of which is manifested in change in temperature of the global ocean.\textsuperscript{36}

An initial indication of the equilibrium sensitivity of Earth’s climate to a radiative perturbation may be gained from the model of the climate system without feedbacks given by Eqs. (1)-(3) in Resource Letter GECC-1, §II. In the absence of feedbacks, that is, neither the planetary co-albedo, $\gamma$, nor the effective emissivity, $\varepsilon$, (Resource Letter GECC-1, §II) is a function of surface temperature, $T_s$, the derivatives in Eq. (4) are

$$\frac{\partial Q}{\partial T_s} = 0; \quad \frac{\partial E}{\partial T_s} = -4\varepsilon\sigma T_s^3 = -\frac{4}{T_s} \varepsilon\sigma T_s^4 = -\frac{4}{T_s} \frac{\gamma J_s}{T_s} = -\frac{\gamma J_s}{T_s}, \quad (8)$$

from which the sensitivity of this no-feedback climate system, denoted by the subscript NF, is
\[ S_{\text{NF}} = \frac{T_s}{\gamma J_S} \]  

This no-feedback sensitivity is the equilibrium sensitivity of a Stefan-Boltzmann–like planet that is gray in both the shortwave (co-albedo \( \gamma = 0.71 \)) and the longwave (emissivity \( \varepsilon = 0.62 \)) but with both grayness quantities held constant is about 0.30 K/(W·m\(^{-2} \)). If the assumptions of this model were accurate for Earth, we would have solved a big part of the science question having to do with climate change. Unfortunately, the problem is much more complicated than this because the actual response of Earth’s climate system to perturbations includes the effects of feedbacks.

Historical understanding of the magnitude of Earth’s actual “equilibrium” climate sensitivity is summarized in Figure 5, which shows best estimates of this quantity and of associated uncertainties going back to the original estimate by Arrhenius\(^{35} \) (Nobel Prize in chemistry for ionic dissociation; activation energy of chemical reactions). Arrhenius obtained his estimate by extensive calculations as a function of latitude and season, taking into account snow/ice cover, relative humidity, cloudiness, and the absorption spectra of water vapor and carbon dioxide. Also shown for reference is the no-feedback sensitivity just calculated, with the date being that of the determination of Stefan’s constant. (There is no indication that Stefan calculated this sensitivity, but he did use his law to obtain a very accurate determination of the temperature of the Sun\(^{39} \))

Despite much work, especially over the past several decades, Earth’s equilibrium sensitivity remains quite uncertain, for a variety of reasons, as summarized in the 2013 IPCC Assessment (Ref. [4], §TFE.6). As the value of this sensitivity has major implications on strategies to constrain global temperature rise due to increases in GHG concentrations, determination of Earth’s climate sensitivity is the “holy grail” of current climate change research.
Figure 5. Historical estimates of Earth’s “equilibrium” climate sensitivity. Sensitivities are given as long-term temperature increase $\Delta T_{2\times}$, left axis, that would result in response to sustained doubling of atmospheric CO$_2$ and in systematic units, K/(W·m$^{-2}$), right axis, with the conversion (forcing due to doubled CO$_2$) taken as 3.7 W·m$^{-2}$. The value denoted Stefan represents the no-feedback sensitivity calculated, Eq. (15), using the Stefan-Boltzmann radiation law but was not actually calculated by Stefan. The value denoted Arrhenius is by direct calculation. $^{37}$ Charney NRC is from an assessment by the U. S. National Research Council.$^{38}$ The several ranges (central 66% of the likelihood distribution) and, for 2007, point and range, in red are best estimates in the several reports of the Intergovernmental Panel on Climate Change, IPCC; also shown for 2013 is central 90% range, dashed.

Attention is called to the two ordinate scales on the graph in Figure 5. For historical reasons that go back to Arrhenius’ 1896 estimate$^{37}$, the sensitivity of the climate system to a radiative perturbation is commonly expressed as long-term temperature increase $\Delta T_{2\times}$ that would result from a sustained doubling of atmospheric CO$_2$, denoted here and elsewhere as the CO$_2$ doubling temperature of the planet. This quantity is also commonly denoted the “equilibrium climate sensitivity” of the planet, but as stressed above the system is not an equilibrium system and the quantity is a temperature change for a specific forcing, not a sensitivity. Expressing the long-term sensitivity as $\Delta T_{2\times}$, denominated on the left-hand axis of the figure, particularizes the sensitivity as the response to CO$_2$ doubling; this particularization raises questions, for example, about possible dependence on the magnitude of CO$_2$ mixing ratio at the initial state and about the response to a perturbation other than that of doubling CO$_2$ (other magnitudes of CO$_2$ perturbation or other kinds of radiative forcings). The systematic unit for sensitivity, K/(W·m$^{-2}$),
as determined in the above calculation for the no-feedback climate sensitivity and denominated on the right-hand axis, has not gained much traction, much as it might seem to be preferred. The conversion between the two quantities is

\[ \Delta T_{2\times} = F_{2\times}S_{eq}, \]

where \( F_{2\times} \) denotes the forcing that would result from doubling atmospheric CO\textsubscript{2}, commonly taken as 3.7 W\cdot m\textsuperscript{-2} (Table 2; Refs [18, 19]). If \( F_{2\times} \) were accurately known, then converting between doubling temperature \( \Delta T_{2\times} \), unit K, and sensitivity \( S_{eq} \), unit K/(W\cdot m\textsuperscript{-2}), would pose no problem. However, the central 90% uncertainty range in the radiative forcing by doubled CO\textsubscript{2}, as given in the most recent (2013) IPCC Assessment, [± 20%] (Ref [4], Technical Summary §TFE4), is substantial.

All of the estimates of Earth’s equilibrium climate sensitivity shown in Figure 5 and virtually all current estimates\textsuperscript{40} exceed the no-feedback sensitivity and are therefore indicative of positive feedback in the climate system. The best estimate and range denoted “NRC” comes from an assessment conducted under the auspices of the National Research Council of the United States (NRC)\textsuperscript{38} that was informed largely by early calculations with GCMs, general circulation models of Earth’s atmosphere. Although remarkable for their time, the models of that time were quite primitive by the standards of today’s climate models. The several ranges and, for the 2007 assessment, best estimate and range, of the successive Assessment Reports of the IPCC are based on considerations of multiple sources of information: GCM calculations, energy balance estimates, and estimates from paleo climates. Also shown in Figure 5 is the estimate of the central 90% of the likelihood distribution for climate sensitivity given in the 2013 Assessment, the lower end of which coincides with the no-feedback sensitivity.
As a summary of the present state of understanding of Earth’s climate sensitivity, Figure 5 might be viewed as “good-news, bad-news.” The good news is that the best estimates given in the several studies and assessments are all about the same. The bad news is that the present estimate of this sensitivity is uncertain to a factor of 3, where the uncertainty range represents the central 66% of the likelihood distribution, approximately ± 1 \( \sigma \). Further bad news is that despite all the intervening research, the estimates of the uncertainty range have not diminished between the 1979 NRC report and the 2013 IPCC report. This uncertainty thus remains quite large relative to the accuracy, say 25%, required to develop emission control strategies to limit the future increase in GMST to a desired value.

An important point pertinent to quantifying past and prospective climate change has to do with the small magnitudes of the changes in radiative fluxes and global temperature relative to the magnitudes of the initial, unperturbed quantities. The observed change in GMST of about 0.8 K (Ref. [4], §2.4) represents a change of about 0.3% relative to the initial 287 K. Even the 2 K increase in GMST that is widely considered a threshold of dangerous anthropogenic interference with the climate\(^{41-44}\) and which is the target maximum increase in temperature specified in the 2015 Paris Agreement of the United Nations Framework Convention on Climate Change\(^{45}\) represents a change in GMST of less than 1%. The challenge to the climate change research community is to gain quantitative understanding of the changes in quantities influencing climate change and the expected response of the system to the accuracy necessary for informed decision making regarding prospective controls on future emissions of climate influencing substances. Such quantitative understanding is essential to answering “what if” questions regarding the consequences of future emissions of climate influencing substances. Determining the increase in
GMST that would result if the mixing ratio of CO$_2$ were to increase to twice its preindustrial value to some desired level of accuracy, say 25%, is a daunting challenge.


35. Bjorn Stevens, private communication.


V. INFERENCES AND IMPLICATIONS

Knowledge of the forcing over the Anthropocene together with Eq. (7b) relating forcing to expected change in global mean surface temperature (GMST) serves as the basis for comparison of expected and observed change in GMST. The observed increase in GMST over the period of globally representative instrumental measurements from the second half of the 19th century to 2011 given by the 2013 IPCC Assessment (Ref. [4], § 2.4.3) was about 0.8 K (https://crudata.uea.ac.uk/cru/data/temperature/HadCRUT4-gl.dat). The best estimate total forcing over the period 1850-2011 is 2.2 W·m⁻² (Figure 2). The present planetary heating rate inferred from increase in the heat content of the planet over time, N, also taken from the 2013
IPCC Assessment (Ref. [4], Box 3.1) is about 0.5 W·m⁻². For the equilibrium climate sensitivity expressed as CO₂ doubling temperature $\Delta T_{2\times}$ taken as the central value of the 2013 IPCC estimate, 3 K, and the forcing of doubled CO₂ taken as 3.7 W·m⁻², the calculated increase in GMST, 1.4 K, is substantially greater than the observed increase. This discrepancy is shown graphically in Figure 6, which shows the increase in GMST calculated by Eq. (7b) with $N = 0.5$ W·m⁻² as a function of total forcing and equilibrium climate sensitivity (expressed as $\Delta T_{2\times}$).

The calculated increase in GMST, 1.4 K, falls well to the right of the observed value $\Delta T = 0.8$ K denoted by the thick contour line. However, the discrepancy is readily resolved within present uncertainties in forcing and climate sensitivity as the contour denoting the observed increase in GMST 0.8 K can be achieved by a wide-ranging mix of forcing and climate sensitivity within their respective uncertainties denoted by the vertical and horizontal pairs of lines. Thus the observed increase in GMST can be achieved for $\Delta T_{2\times}$ at the low end of the range given by the 2013 IPCC Assessment, 1.5 K, if the total forcing is 2.5 W·m⁻² or, alternatively, for $\Delta T_{2\times}$ at the high end of the IPCC range, 4.5 K, if the total forcing is 1.1 W·m⁻², or anywhere in between.

Indeed, if the forcing is at the high end of the IPCC range, 3.3 W·m⁻², $\Delta T_{2\times}$ must be 1 K, a very low sensitivity, essentially equal to the no-feedback sensitivity (Figure 5). On the other hand, if the forcing is at the low end of the uncertainty range, 1.3 W·m⁻², then $\Delta T_{2\times}$ must be 5 K, quite a high sensitivity. This analysis shows that the present uncertainty in total forcing precludes observationally constraining climate sensitivity, and vice versa.
Figure 6. Dependence of expected increase in global mean surface temperature (accounting for global heating rate taken as 0.5 W·m⁻²; Ref. 5, Box 3.1) on total forcing over the Anthropocene, abscissa, and equilibrium climate sensitivity expressed as long-term temperature increase \( \Delta T_{2x} \) that would result in response to sustained doubling of atmospheric CO₂, ordinate. Vertical black lines denote central 90% of the likelihood distribution of forcing and horizontal lines denote central 66% of the likelihood distribution of climate sensitivity, both as given by the 2013 IPCC Assessment. Bold contour at 0.8 K denotes approximate value of observed increase in GMST. Points denote expected increase in GMST due to IPCC best estimate total forcing over the Anthropocene, 2.29 W·m⁻² and by long-lived greenhouse gases only, 2.82 W·m⁻² (Figure 2).

This situation has major implications regarding developing strategies to limit increase in global temperature. Importantly, reducing fossil fuel combustion to limit CO₂ emissions would also reduce emissions of aerosols and precursor gases. As the atmospheric residence time of these aerosols is quite short (weeks) relative to that of CO₂ and other greenhouse gases (decades to centuries), reducing fossil fuel emissions would thus result in a rapid decrease of the negative aerosol forcing. In turn, total forcing and GMST would increase. Under this situation, the total forcing would increase, in the limit, to the present forcing by long-lived GHGs, 2.8 W·m⁻², denoted “GHG” in Figure 6. For climate sensitivity at its best estimate value, the resulting temperature increase relative to preindustrial would be 1.9 K. Within present uncertainty in climate sensitivity, as given by the 2013 IPCC Assessment, the temperature increase above preindustrial to which the planet is committed on account of long-lived GHGs now in the
atmosphere would range from the 1 K already realized to, at the high end of the sensitivity range, as great as 2.8 K [Ref. 46]. The latter substantially exceeds the widely accepted 2 K threshold for dangerous anthropogenic interference with the climate system (§IV). The possibility that the negative forcing by anthropogenic aerosols has prevented the increase in GMST that would otherwise have been much greater has been characterized as a “Faustian bargain.”47 To forestall the full impact of the incremental GHGs, the aerosol offset or some other form of geoengineering would need to be maintained long after the benefit gained from combustion of fossil fuels has been enjoyed.

With respect to societal decision making on future emissions of CO$_2$ and other greenhouse gases the consequences of the present uncertainties are quite discomfitting. If climate sensitivity is at the low end of the uncertainty range, then perhaps there is time for an orderly transition from reliance on fossil fuels as the primary source of the world’s energy to alternative sources of energy. If, however, climate sensitivity is at the high end of that range, then future emissions will add to the already committed increase in GMST that is already well beyond the 2-K threshold that is considered to represent the onset of dangerous interference with Earth’s climate.


VI. SUMMARY

This Resource Letter has reviewed the so-called greenhouse effect of trace species in the atmosphere, water vapor, CO₂ and other polyatomic molecules that absorb and radiate in the thermal infrared, and clouds, calling attention to pertinent primary and secondary literature. The greenhouse effect is a key element of Earth’s climate system responsible for a global mean surface temperature, average roughly 287 K, much greater than the radiative temperature at the top of the atmosphere, 255 K. Because of human activities, largely fossil fuel combustion, but including also deforestation, agricultural activities, and other industrial activities, amounts of long-lived greenhouse gases in the atmosphere, have increased substantially relative to their preindustrial values. The radiative forcing of the incremental greenhouse gases over the Anthropocene to date, calculation of which rests rather confidently on well-understood molecular properties and well characterized properties of the climate system, is of order 1% relative to top-of-atmosphere fluxes, small relative to spatial and temporal variability of these fluxes on a variety of scales. However even a 1% increase in global mean surface temperature, ~3 K, would be of great consequence to human society and natural ecosystems. The uncertainties have enormous implications regarding interpretation of increase in GMST to date, committed future increases in GMST, and policy regarding future emissions of long-lived greenhouse gases.
The physics of the greenhouse effect and more broadly of Earth’s climate and climate change is a research field that has attracted many good physicists and continues to present important first-order challenges and research opportunities.

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Stephen E. Schwartz

SN1. Calculation of radiative forcing by an incremental greenhouse gas

As noted in Resource Letter GECC-1 §IV, the top of the atmosphere (TOA) greenhouse effect consists of two components: the decrease in outgoing longwave radiation due to absorption of irradiance by the greenhouse gas (GHG) and the increase in this radiative flux due to emission by the gas. Here expressions are developed for these effects for an optically thin absorbing gas and are applied to evaluate the forcing by CCl₂F₂ as a readily developed example. CCl₂F₂ (dichlorodifluoromethane) is a synthetic gaseous compound that had been widely manufactured and used as a refrigerant and for other purposes until it became recognized by Molina and Rowland\textsuperscript{51} (Nobel Prize in chemistry, 1995) that chlorine atoms and molecular free radicals, produced in the stratosphere through photolysis by ultraviolet radiation, catalyzed the destruction of ozone, with many attendant consequences. Manufacture of this compound (and also of CCl₃F) was largely banned under the Montreal Protocols,\textsuperscript{52} but because of their long residence times these compounds are persistent in the atmosphere, and their amounts are only slowly decreasing. As polyatomic molecules, these compounds exhibit numerous radiatively active vibrational transitions in the thermal infrared. Importantly, several of these transitions occur in the so-called “window region” of the infrared spectrum (\(\sim 800 – 1000\) and \(\sim 1200 – 1300\) cm\(^{-1}\); Resource Letter GECC-1, Figure 4) in which there is little absorption by the major infrared absorbing gases, water vapor or carbon dioxide, and in which, therefore, infrared radiation from the surface is transmitted to space, relatively unimpeded. This situation greatly enhances the greenhouse effect of these gases relative to what it would be in other spectral regions, and greatly simplifies calculation of the magnitude of this effect. An example of such a calculation is presented here.
For any absorption line or band, the rate of absorption of radiant energy is calculated as the product of the absorption coefficient times the incident irradiance integrated over the frequency range, $d\nu$, for which there is appreciable absorption. For the incident irradiance at frequency, $\nu$, given by the Planck function, $J_{\text{sfc},\nu}(\nu, T_{\text{sfc}})$, at the surface temperature, $T_{\text{sfc}}$, the volumetric rate of absorption of upwelling radiant energy at a given altitude, $z$, is

$$P_{\text{abs}}(\nu, z) = n_i(z) \int J_{\text{sfc},\nu}(\nu', T_{\text{sfc}}) \sigma(\nu', z) d\nu',$$  \hspace{1cm} (S1.1)

where $n_i(z)$ is the number concentration of absorbing molecules of species $X_i$ and where $\sigma(\nu, z)$ is the absorption cross section of the molecule. The absorption cross section is peaked at the center frequency, $\nu$, of the transition; it is weakly dependent on altitude because of pressure-dependent collisional broadening of the transition and temperature-dependent Doppler broadening. The integral is taken over frequencies at which there is appreciable absorption cross section. The absorbed power has the unit, $\text{W} \cdot \text{m}^{-3}$; the Planck function is expressed in the unit $\text{W} \cdot \text{m}^{-2} \cdot \text{Hz}^{-1}$, with frequency, $\nu$, in Hz. For $n_i(z)$ in molec $\cdot$ m$^{-3}$, $\sigma(\nu, z)$ has the unit m$^2$ $\cdot$ molec$^{-1}$.

As the Planck function is slowly varying with frequency over a given absorption band, it may be pulled out of the integral,

$$P_{\text{abs}}(\nu, z) = J_{\text{sfc},\nu}(\nu, T_{\text{sfc}}) n_i(z) \int \sigma(\nu', z) d\nu'.$$  \hspace{1cm} (S1.2)

Further, denoting the integral $\int \sigma(\nu', z) d\nu' = S_\nu(\nu, z)$, yields

$$P_{\text{abs}}(\nu, z) = J_{\text{sfc},\nu}(\nu, T_{\text{sfc}}) S_\nu(\nu, z) n_i(z).$$  \hspace{1cm} (S1.3)

To good approximation the integral over an absorption line or band, the band strength or integrated absorption coefficient, $S_\nu = \int \sigma(\nu) d\nu$, is independent of pressure and temperature broadening, and hence the altitude dependence of the band strength may be neglected to first order. The integrated absorption
coefficient, $S_\nu$, is related to the integrated absorption coefficient determined in laboratory measurements and commonly reported as $S_\nu = \int \alpha(\tilde{\nu})d\tilde{\nu}$ having unit cm·molec$^{-1} = $ cm$^2$·molec$^{-1}$·cm$^{-1}$, where $\tilde{\nu}$ denotes wavenumber (unit, cm$^{-1}$) as $S_\nu = S_\nu \times 10^{-4}$ [m$^2$·cm$^{-2}$] $c$[cm·s$^{-1}$] = $2.98 \times 10^6$ [m$^2$·cm$^{-1}$·s$^{-1}$] $S_\nu$.

The total absorption of upwelling irradiance, per area, in an atmospheric column is given by the integral of Eq. (S1.3) over the height of the column; this is the top-of-atmosphere absorption forcing, the decrease in emitted irradiance at the top of the atmosphere due to absorption by the gas,

$$F_{abs}^{toa}(\nu) = J_{sfc,\nu}(\nu, T_{sfc}) S_\nu \int n_i(z)dz ,$$ (S1.4)

with unit W·m$^{-2}$. For a well-mixed gas, $n_i(z) = x_i n_{air}(z)$, where $x_i$ is the mixing ratio (mol/mol air), and $n_{air}$ is the number concentration of air molecules, so that the integral over height can be replaced by an integral involving the concentration of air molecules.

$$F_{abs}^{toa}(\nu) = J_{sfc,\nu}(\nu, T_{sfc}) S_\nu x_i \int n_{air}(z)dz .$$ (S1.5)

The integral in Eq. (S1.5) is the number of molecules of air, $N_{air}$, per square meter in the atmospheric column,

$$\int n_{air}(z)dz = N_{air} = \frac{P}{m_{air}g} = N_{air}^0 P_{sfc} ;$$ (S1.6)

here $m_{air}$ is the mass of a molecule of air calculated for the average molecular weight (molar mass) of air, 0.029 kg·mol$^{-1}$, and $P_{sfc}$ is the surface pressure in systematic units (Pa); for standard pressure $P_{sfc}^0$ taken as 1 bar (10$^5$ Pa), $N_{air}^0 = 2.12 \times 10^{29}$ molec·m$^{-2}$. This yields the rather simple expression for the top of the atmosphere (TOA) absorption forcing in terms of surface pressure in bar

$$F_{abs}^{toa}(\nu) = J_{sfc,\nu}(\nu, T_{sfc}) S_\nu x_i N_{air}^0 P_{sfc}^0 / bar .$$ (S1.7)
As noted in Resource Letter GECC-1 §IV, it follows from Kirchoff’s law that if an infrared-active gas were in an isothermal system at the temperature of the surface, the power emitted by the gas (in each direction, up and down) would be equal to the power absorbed. However, the gas being at altitude higher than the surface, it is also at lower temperature; this dependence of temperature on altitude strongly influences the emitted power. The reduction in emitted power is a consequence of the lower population of molecules in the excited state than would be present at equilibrium at the temperature of the surface. This temperature dependence is given by the Boltzmann factor for the molecular vibrational state responsible for the emission. Hence, the amount by which the rate of emission is less than that if the system were isothermal is given by the factor \( \exp[-h\nu(T(z)^{-1}-T_{sfc}^{-1})/k] \), where \( T(z) \) is the local atmospheric temperature, a function of altitude, and where \( h \) and \( k \) are the Planck and Boltzmann constants, respectively. Thus, at a given altitude, \( z \), the volumetric rate of emission is

\[
P_{\text{emit}}(v,z) = P_{\text{emit}}^+(v,z) = J_{\text{sfc},v}(v,T_{\text{sfc}}) x_i \ S_v \ n_{\text{air}}(z)e^{-h\nu(T(z)^{-1}-T_{\text{sfc}}^{-1})/k}.
\]  

(S1.8)

The Boltzmann factor governing the emitted power is referenced to the surface temperature because the Planck emission factor is that for the surface temperature. The local volumetric heating rate is the local difference between absorption and emission

\[
P_{\text{net}}(z) = P_{\text{abs}}^+(z) - P_{\text{emit}}^+(z) = P_{\text{abs}}^-(z) - P_{\text{emit}}^+(z) = P_{\text{abs}}^+(z) - 2P_{\text{emit}}^+(z).
\]  

(S1.9)

There is no contribution from downwelling irradiance because the gas is treated as optically thin and because the atmosphere is otherwise non-absorbing in the wavelength range under consideration here, the window region. The cumulative change in upwelling irradiance due to the presence of the infrared-active gas from the surface to altitude \( z \) is given by the vertical integral of the difference in absorbed and upward emitted power; this is the net forcing of upwelling irradiance due to a specific absorption band of species \( i \) as a function of altitude,
\[ F_{\text{net}}(\nu, z) = \int_0^z \left[ P_{\text{abs}}(\nu, z') - P_{\text{emit}}(\nu, z') \right] \, dz' = J_{\nu, \text{sfc}}(\nu, T_{\text{sfc}}) \, S_{\nu} x_i \int_0^z n_{\text{air}}(z')[1 - e^{-h\nu(T(z')^{-1} - T_{\text{sfc}}^{-1})}] \, dz' . \]  

(S1.10)

The net TOA forcing is given by the integral over the entire height of the vertical column.

Equation (S1.10) permits evaluation of forcing that would result from a given transition of a given infrared-active gas by numerical integration for a particular atmospheric temperature profile, with the requirements that the scene be cloud-free and that the gas be optically thin and that the transition occur in a region of the spectrum that is not obscured by other infrared-active gases. The results of the calculation are illustrated in Figure S1 for the 925 cm\(^{-1}\) band of CCl\(_2\)F\(_2\), which is in the window region of the thermal infrared, Resource Letter GECC-1, Figure 4. The band strength, based on laboratory measurements,\(^{S3}\) \(S_{\nu} = 4.68 \times 10^{-17} \text{ cm \cdot molec}^{-1}\), corresponds to \(S_{\nu} = 1.40 \times 10^{-10} \text{ m}^2 \cdot \text{molec}^{-1} \cdot \text{Hz}\). The atmospheric mixing ratio of the gas \(x_i\) is 0.55 ppb (5.5 \(\times\) 10\(^{-10}\)), Resource Letter GECC-2, Table 2. The vertical temperature profile is taken as that in Resource Letter GECC-1, Figure 5. The resulting heating rate and forcing profiles, Figure S1, show the heating rate negative (i.e., cooling) at low altitude, warming at high altitude, with the absorption contribution to forcing of upwelling irradiance dominant, offset by roughly half by the emission forcing. The net TOA forcing for this vibrational band is 0.067 W\(\cdot\)m\(^{-2}\). The sum of forcings over the several vibrational bands of the molecule, calculated in the same way, gives a total TOA forcing of 0.17 W\(\cdot\)m\(^{-2}\). Although this forcing is in close agreement with the value given in Resource Letter GECC-2, Table 2, the close agreement must be viewed as somewhat fortuitous. The calculation, carried out for an equatorial temperature profile, neglects spatial and temporal variation of surface and atmospheric temperature and as the actual forcing would be expected to be less than that calculated by the simple approach because of obscuration of the forcing by clouds. Nonetheless, this example illustrates the approach to calculation of radiative forcing by incremental
greenhouse gases and the vertical profiles of heating and cooling by absorption and emission of radiation by a greenhouse gas.

![Graph](fs02 RompsHeatingRateForcing925.eps)

**Figure S1.** Vertical profiles, *left*, of heating (or cooling, negative) rate and, *right*, of cumulative contribution to TOA forcing by the 925 cm\(^{-1}\) band of CCl\(_2\)F\(_2\) (positive forcing denotes increase in global heat content). Red curves denote absorption of irradiance emitted at the surface by the infrared-active gas; blue curves denote emission by the gas, and green curves denote the net effect.

Further examples of such calculations are given in Refs. [S4] and [S5].


SN2. Can forcings from increases in GHGs be measured?

In principle, as the amount of a greenhouse gas (GHG) is increased in the atmosphere, it should be possible to measure the resultant change in irradiance, at the top of the atmosphere or at the surface. Working against this, however, would be the effects of variation of temperature and water vapor, on a variety of time scales that would exert large fluctuations in infrared power that would overwhelm any change due to the incremental amount of GHG. Moreover, as a consequence of the presence of the incremental amount of GHG, there would, over time, be an increase in surface temperature and resultant increase in thermal infrared emitted at the surface. This would offset the forcing by the incremental GHG as follows; because of the increase in surface temperature there would likely be an increase in the temperature and, in turn, of the amount of water in the atmospheric column that would, through the greenhouse effect of water vapor amplify the effect of the initial perturbation (positive feedback). As well, there would likely be changes in the amount and distribution of clouds that would contribute further, with sign not yet known, to change in net power at the top of the atmosphere (TOA). Given this situation would it be possible to measure the forcing due to the increment of GHG over a period of time? If measurements were limited to total broadband thermal infrared, the answer would be a resounding no. However, because the several greenhouse gases exhibit unique spectral features, the possibility arises that it might be possible to discern and quantify the changes attributable to increases in individual GHGs


from examination of changes in the spectral distribution of the infrared radiation emitted at the TOA. Such an examination\textsuperscript{86} using 1970 measurements from Nimbus-4 (\textit{cf}. Resource Letter GECC-1, \textbf{Figure 4}) and similar measurements from 1997 reported decreases in emitted radiance in absorption bands of CO\textsubscript{2}, CH\textsubscript{4}, O\textsubscript{3}, chlorofluorocarbons, and water vapor that were attributed to increases in these gases over this period.

As the greenhouse effect is manifested at the surface as well as at the TOA, secular increases might similarly be expected in downwelling infrared radiance at the surface. Using spectrally resolved measurements at the surface in Oklahoma and at the North Slope of Alaska, Feldman et al.\textsuperscript{87} examined changes in downwelling irradiance over 2000-2010, during which period CO\textsubscript{2} mixing ratio increased by 22 ppm. Although the observed trends in downwelling radiance were dominated by trends in humidity and temperature, it was possible to account for those trends using measured temperature and humidity profiles as input to a radiation transfer model. This permitted quantification of the increase in downwelling radiance in the wings of the CO\textsubscript{2} absorption bands, where the transitions are not saturated, consistent with radiation transfer model calculations for the increase in CO\textsubscript{2} over the period. The increase in surface forcing due to the increase in CO\textsubscript{2}, calculated as the spectrally integrated increase in downwelling irradiance, was found at both sites to be $0.2 \pm 0.06 \ (1\sigma) \ W \cdot m^{-2} \cdot \text{decade}^{-1}$, equivalent to a dependence on CO\textsubscript{2} mixing ratio of $0.010 \pm 0.003 \ W \cdot m^{-2} \cdot ppm^{-1}$. Radiation transfer model calculations indicated that the increase in the TOA forcing over the period, although somewhat greater than at the surface, was still considerably less than the value in Resource Letter GECC-2, Table 2, 0.016 $W \cdot m^{-2} \cdot ppm^{-1}$.

Returning to the question of whether the GHG forcings can be measured, the answer would thus seem to be a qualified yes, but only with use of radiation transfer models to account for other concurrent changes.
in the temperature and humidity structure that can dominate changes in the longwave irradiance. Nonetheless, the observed changes in the spectral distribution of the radiation lend strong confirmation to the role of incremental GHGs in increasing the greenhouse effect of the planet and to the understanding of the governing processes that is represented in radiation transfer models.


**SN3. The budget and adjustment time of incremental atmospheric CO₂**

Evaluation of the adjustment time of incremental CO₂ (*i.e.*, in excess of preindustrial) is illustrated in **Figure S2** (see also Refs. [S10] and [S11]). The amount of excess CO₂ in the global atmosphere (expressed as mass of carbon, as is conventional) in panel (a) is based on contemporaneous measurements, Figure 1; 1 Pg = 1 × 10¹⁵ g. Panel (b) shows the budget of sources and sinks of CO₂ into/out of the atmosphere. The rate of emissions, at present and historically, is known fairly accurately from inventories of fossil fuel combustion and deforestation; in the aggregate current emissions are about 10 Pg·yr⁻¹. The principal source of incremental CO₂ at present is fossil fuel combustion, with a further contribution from calcination of limestone in the manufacture of cement; land use change (net of deforestation minus afforestation) contributes about 15% to present global anthropogenic CO₂ emissions. This annual increment of CO₂ in the atmospheric reservoir is diminished by removal – net uptake by the terrestrial biosphere and transport from the upper ocean to the deep ocean – the rate of
which is inferred from observations as the difference between emissions and the annual increase of atmospheric CO\textsubscript{2}. The ocean sink was estimated using the average of four global ocean biogeochemistry models forced by observed atmospheric conditions and CO\textsubscript{2} mixing ratio, with the land sink estimated as the residual. The substantial fluctuations in the total annual sink and in the inferred land sink, panel (c), are not confidently understood. Proportionality of the sink rate to the excess atmospheric CO\textsubscript{2}, panel (d), is consistent with removal of excess CO\textsubscript{2} with a single time constant of about 40-50 years, evaluated as the inverse of the slope and is consistent with evaluation of this quantity as the ratio of excess atmospheric CO\textsubscript{2} to the sink rate, panel (e). Such an adjustment time, much shorter than commonly given for this quantity, 150 – 500 yr [e.g., Refs. S12-S14] is consistent with other empirically based estimates [Refs. S15 and S16].

It is generally thought (e.g., Refs. [S11-S14]) that well into the future (time scale of millennia), as the deep ocean becomes increasingly saturated in excess CO\textsubscript{2}, the effective rate coefficient for removal of excess CO\textsubscript{2} from the atmosphere-upper-ocean compartment would decrease. Ultimately, at long time (greater than 500 years following a cessation of emission of CO\textsubscript{2} into the atmosphere), the decay of excess atmospheric CO\textsubscript{2} would no longer be characterized by a single time constant but would plateau out at a non-zero value. This can be estimated, based on the solubility of CO\textsubscript{2} and the volume of the deep ocean, as about 20% of the total incremental CO\textsubscript{2} added to the system.
**Figure S2.** (a) Annual average global excess atmospheric CO$_2$ based on measurements at Mauna Loa, Hawaii; the unit Pg denotes petagrams (10$^{15}$ g) of carbon. (b) Annual sources and sinks of atmospheric CO$_2$, 1959-2011 (data from Ref. [S8]; http://cdiac.ornl.gov/GCP/carbonbudget/2013/). Sources (positive) and sinks (negative) are shown in summation to categories closer to the x-axis. Fossil fuel combustion and cement production are from historical records. Emissions from land use change (net of deforestation and afforestation; Ref. [S9]), are uncertain by 50%. Atmospheric growth is from measurements at Mauna Loa (1959-1980) or global average (1980-2010). (c) CO$_2$ removal rate (annual, points, and 3-year smoothing, blue curve) calculated as difference between annual emissions and measured annual atmospheric increment. (d) CO$_2$ adjustment time evaluated as excess CO$_2$ divided by smoothed sink rate. (e) Sink rate vs. excess CO$_2$ and linear fits, forced through origin, for entire data set and for first and second halves, indicated by dotted and dashed lines, respectively, with corresponding adjustment times.


Supplementary Notes to “Greenhouse Effect and Climate Change-2,” Stephen Schwartz


**SN4. Treating global temperature response as linear in the perturbation**

How much error results from considering $\Delta E$ linear in $\Delta T_s$? Consider a black body emitting at the global mean surface temperature of Earth, $T_s$. The emitted flux is

$$E_{bb}' = \sigma T_s^4$$

(S4.1)

where for convenience here the emitted flux is taken here as positive outward. The temperature is increased by $\Delta T_s$, resulting in an increase in the magnitude of the emitted flux $\Delta E_{bb}$. The new emitted flux is

$$E_{bb}' = \sigma (T_s + \Delta T_s)^4$$

(S4.2)

which yields, to second order,

$$\Delta E_{bb} = 4\sigma T_s^3 \Delta T_s [1 + \frac{3}{2} \frac{\Delta T_s}{T_s} + \cdots]$$

(S4.3)

Consider a 1% perturbation in $T_s$, about a 3 K increase over the present 287 K. The second term of the quantity in brackets is about 0.015. The take-away message here is that the change in climate that is under consideration is a small perturbation from the initial state, so that the error that results from treating the response as linear is small, at least at the present level of understanding, justifying the treatment of climate system response as linear.