

Short-Lived Climate Pollution

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Abstract

Although carbon dioxide emissions are by far the most important mediator of anthropogenic climate disruption, a number of shorter-lived substances with atmospheric lifetimes of under a few decades also contribute significantly to the radiative forcing that drives climate change. In recent years, the argument that early and aggressive mitigation of the emission of these substances or their precursors forms an essential part of any climate protection strategy has gained a considerable following. There is often an implication that such control can in some way make up for the current inaction on carbon dioxide emissions. The prime targets for mitigation, known collectively as short-lived climate pollution (SLCP), are methane, hydrofluorocarbons, black carbon, and ozone. A re-examination of the issues shows that the benefits of early SLCP mitigation have been greatly exaggerated, largely because of inadequacies in the methodologies used to compare the climate effects of short-lived substances with those of CO₂, which causes nearly irreversible climate change persisting millennia after emissions cease. Eventual mitigation of SLCP can make a useful contribution to climate protection, but there is little to be gained by implementing SLCP mitigation before stringent carbon dioxide controls are in place and have caused annual emissions to approach zero. Any earlier implementation of SLCP mitigation that substitutes to any significant extent for carbon dioxide mitigation will lead to a climate irreversibly warmer than will a strategy with delayed SLCP mitigation. SLCP mitigation does not buy time for implementation of stringent controls on CO₂ emissions.

Atmospheric lifetime: the characteristic time during which a substance added to the atmosphere remains in the atmosphere

Short-lived climate pollution (SLCP): substances added to the atmosphere by human activities, that alter the climate and that persist in the atmosphere for under a few decades

1. INTRODUCTION

Human activities result in the emission of a number of substances that alter the composition of Earth's atmosphere and in consequence alter the climate. The most important of these is carbon dioxide (CO₂), emitted primarily as a result of burning fossil fuels such as coal, oil, or natural gas and secondarily as a result of land use changes, notably deforestation. In addition, agricultural and industrial activities emit a range of other greenhouse gases, which, like CO₂, lead to a warmer climate by affecting the efficiency with which the planet cools by infrared emission to space. Most of these gases have atmospheric lifetimes of a decade or more, which allows sufficient time for their concentrations to become nearly uniform throughout the troposphere. Combustion also leads to the production of a wide variety of particulate atmospheric contaminants, with residence times typically from days to weeks. These include particulates with primarily a cooling effect through increasing the reflection of solar radiation (sulfate aerosols and secondary organic aerosols) and those with primarily a warming effect through increasing absorption of solar radiation (primarily black carbon). These are too short lived to be well mixed, and they lead to spatially and seasonally inhomogeneous climate forcing. Finally, atmospheric chemistry acting on certain anthropogenic emissions can produce extremely short-lived greenhouse gases in the troposphere—notably ozone—which act similarly to other greenhouse gases but yield very inhomogeneous climate forcing. Particulates and ozone have important health and agricultural effects, which tend to make their elimination necessary regardless of whether such actions help or hinder climate protection.

In this review, we use the term short-lived climate pollution (SLCP) to refer to climate-altering emissions such as methane or black carbon with atmospheric lifetimes of under a few decades. CO₂, in contrast, is persistent enough that the climate remains locked in a significantly hotter state for thousands of years after the cessation of emissions (Section 2). The prime examples of SLCP, and the ones that have received the most attention as targets for mitigation, are methane (CH₄, with an atmospheric lifetime of about 12 years), hydrofluorocarbons (HFCs, with lifetimes up to a few decades), tropospheric ozone (with lifetime up to a month), and black carbon (with an atmospheric lifetime on the order of days and persistence in snow that could extend to several months). Atmospheric contaminants that cause surface cooling, such as sulfate and secondary organic aerosols, are not typically categorized as SLCP, but because human activities produce a mix of warming and cooling aerosols that often cannot be controlled independently, the cooling aerosols must inevitably be considered in analyzing the net effect of mitigation strategies.

There is a clear need to understand the full range of climate-relevant emissions, including SLCP, in order to make sense of what is happening to our climate and test climate models against observations. The growing body of work on SLCP has made a valuable contribution to that effort. But beyond that scientific goal, study of SLCP has given rise to what can only be described as a movement, based on the premise that immediate and aggressive SLCP mitigation would make an important, and perhaps even essential, contribution to climate protection. This point of view has found expression in the scientific literature (Jacobson 2002, Penner et al. 2010, Ramanathan & Xu 2010), national and international assessment reports (UNEP 2011, US EPA 2012), the political science literature (Victor et al. 2012), and the popular press. The SLCP movement has led to the formation of a United Nations entity, the Climate and Clean Air Coalition, whose aim is to promote and implement early mitigation of SLCP (<http://www.unep.org/ccac/>). The work of this coalition has been endorsed by the US State Department (US State Dept. 2012). Several existing governmental and nongovernmental organizations have also promoted early action on SLCP mitigation, including the World Bank (World Bank 2013) and the Arctic Council (DeAngelo 2011). Early SLCP mitigation has been enshrined in US President Obama's

Climate Action Plan (White House 2013). To some extent, the interest in SLCP mitigation arises from frustration over the world's evident inability to come to grips with the problem of CO₂ emissions, which are universally agreed to be the primary threat to the climate, even among supporters of SLCP mitigation. Indeed, the rationale for early SLCP mitigation is sometimes expressed as “buying time” to put CO₂ mitigation into place.

However, a closer examination of the issues, as is reviewed in this article, reveals that much of the current body of work on SLCP has fostered a greatly exaggerated impression of the value of early SLCP mitigation. *Eventual* abatement of SLCP can help reduce peak warming, but the contribution is significant in comparison with the larger CO₂ effect only in a context in which cumulative CO₂ emissions during the time it takes to bring emissions to zero are kept low. Even then, the chief benefits of SLCP abatement are fully realized even if the abatement is delayed until annual CO₂ emissions are declining and are on track to approach zero. There is little to be gained by implementing SLCP abatement measures earlier, and much to be lost if early SLCP abatement to any significant extent displaces CO₂ abatement that would otherwise take place. Early SLCP mitigation does not in any way make up for current inaction on CO₂ mitigation and does not make it any easier to meet climate protection targets through later action on CO₂; it does not “buy time.”

There are a variety of factors that have gone into creating the current widespread misconceptions about the value of early SLCP abatement. These include restriction of the analysis to an overly short time frame, failure to consider strategies involving delayed SLCP abatement, unrealistic assumptions about the amount of SLCP abatement that can be obtained without displacing CO₂ abatement, and insufficient consideration of the amount of SLCP abatement one gets as an automatic cobenefit of CO₂ abatement. For black carbon, there is the additional factor that recent work indicates that the reflecting aerosol precursors committed with most black carbon sources are likely to nullify the warming effects of black carbon. These shortcomings have been compounded by the eagerness of policymakers to appear to be taking some sort of action on climate without having to confront the formidable obstacles to CO₂ mitigation. Even though most analyses of SLCP mitigation have portrayed it as an adjunct to, rather than substitute for, CO₂ mitigation, the message has been lost that SLCP mitigation is essentially useless in the absence of very stringent and immediate measures to restrict CO₂ emissions.

While the scientific literature has so far been dominated by works exploring the short-term benefits of early SLCP mitigation, a few studies have begun to appear that question the value of these benefits and the analyses by which the benefits were assessed (Riahi et al. 2007, NRC 2011, Solomon et al. 2012, Bowerman et al. 2013, Shoemaker & Schrag 2013, Allen & Stocker 2014, Smith & Mizrahi 2014). In this review, we synthesize and extend the results of these studies, in the hopes that this will guide policymakers toward a better understanding of the proper role of SLCP abatement in climate protection strategies. SLCP abatement has value, but it must be distinctly and strictly subordinate to measures aimed at limiting the cumulative amount of CO₂ emitted over the time it takes to bring anthropogenic CO₂ emissions to zero. In fact, a consensus seems to be emerging, embracing both critics and proponents of SLCP abatement, that the value of SLCP abatement can be realized only if such abatement does not interfere to any appreciable extent with efforts to reduce CO₂ emissions, and in particular that no amount of SLCP abatement justifies further delays in CO₂ abatement (Shoemaker et al. 2013).

The core of our argument is quite simple, and is well illustrated by a comparison of the effects of CO₂ versus methane emissions. Each year we add nearly 10 gigatonnes of carbon (GtC) to the atmosphere in the form of CO₂, with the rate steadily increasing; each 10-Gt increment ratchets up Earth's temperature irreversibly by another 0.02°C (Section 2). In 10 years, that adds up to 0.2°C, in a hundred years to 2°C, and inexorably upward so long as emissions continue. Each year

Gigatonnes carbon (GtC): 1 GtC = 3.67 GtCO₂; 1 Gt = 10⁹ metric tonnes = 10¹² kg

we add 0.3 Gt of methane to the atmosphere, and this causes a peak warming of 0.014°C that largely dissipates after a decade (see the **Supplemental Material**; follow the **Supplemental Material link** in the online version of this article or at <http://www.annualreviews.org>). It is simple common sense that we should treat the cause of an irreversible harm first, and defer action on the reversible harm, because action on methane and other short-lived substances gives nearly the same climate protection benefit even if it is delayed by a century; the same cannot be said for delaying CO_2 mitigation. Detailed analysis bears out this reasoning.

The unique features of CO_2 are discussed in Section 2, and the climate effects of other well-mixed greenhouse gases are discussed in Section 3, with a particular emphasis on the way greenhouse gas lifetime affects the translation of emissions into climate change. Part of the misconception about the value of early SLCP mitigation comes from extensive use of a particular emissions metric known as global warming potential (*GWP*) in policy analysis (IEA 2009, Velders et al. 2009, UNEP 2011, among many others). *GWP* forms the basis of most proposed multigas emissions trading schemes and was central to the multigas approach taken in the Kyoto Protocol. *GWP* is a deeply flawed metric, as shown in Section 3.4.1, and gives incorrect policy guidance when used to compare short-lived climate forcing agents with CO_2 . In fact, all emission-equivalent metrics fail in one way or another when used to aggregate short-lived forcing with CO_2 and other gases with millennial effects, so that if a metric is to be used, it must be formulated transparently in light of the specific policy goals it is meant to implement, and it must be changed if those goals change. A brief review of black carbon effects is given in Section 4. An analysis of the consequences of various climate protection strategies involving SLCP mitigation is presented in Section 5; the effects of tropospheric ozone are incorporated in this analysis, though we do not attempt a detailed review of ozone forcing. Our conclusions are summarized in Section 6.

2. WHAT MAKES CO_2 SPECIAL AS A CLIMATE FORCING AGENT?

CO_2 is unique among the well-mixed greenhouse gases in that it is removed from the atmosphere at Earth's surface by a number of processes with disparate timescales ranging from decades to millennia, and that the removal processes have a strong nonlinear dependence on the amount of CO_2 added to the Earth system (Keeling & Bacastow 1977, Walker & Kasting 1992, Archer et al. 1997, Archer et al. 2009). The multiple timescales for CO_2 removal are illustrated in **Figure 1**, which shows the evolution of atmospheric CO_2 concentration following the instantaneous addition of 2,100 GtC to the atmosphere in the form of CO_2 . This is sufficient to raise the concentration from its preindustrial value to 1,250 ppm, but the concentration declines to 1,000 ppm within 20 years, as a result of uptake by the shallow-ocean mixed layer and the land biosphere. But then, the decay slows down as these sinks saturate, and as a result the concentration falls only to a bit under 900 ppm at the end of 100 years. At that point, the main pacemaker of further removal is mixing with the deep ocean, which brings the concentration down to 675 ppm in the next 900 years. At the end of this period, the ocean has taken up as much CO_2 as it can, consistently with the acidification caused by that uptake. Over the next 10,000 years or so the acidity is partially buffered by dissolution of seafloor carbonates and influx of land carbonates from rivers, allowing the ocean to take up a bit more CO_2 . At the end of this process, the concentration is still at 500 ppm, and it will take well over 100,000 years more for the slow processes that react CO_2 with silicate rocks and bind it up in carbonate rocks to restore CO_2 to its preindustrial value. For an elementary exposition of relevant aspects of the carbon cycle, see Pierrehumbert (2010, ch. 8).

When CO_2 dissolves in the ocean, it is repartitioned into carbonate and bicarbonate, and chemical equilibrium equations describing the partitioning play a controlling role in determining how much carbon the ocean can take up. The carbonate equilibrium equations are nonlinear, and

ppm: parts per million
by count of molecules

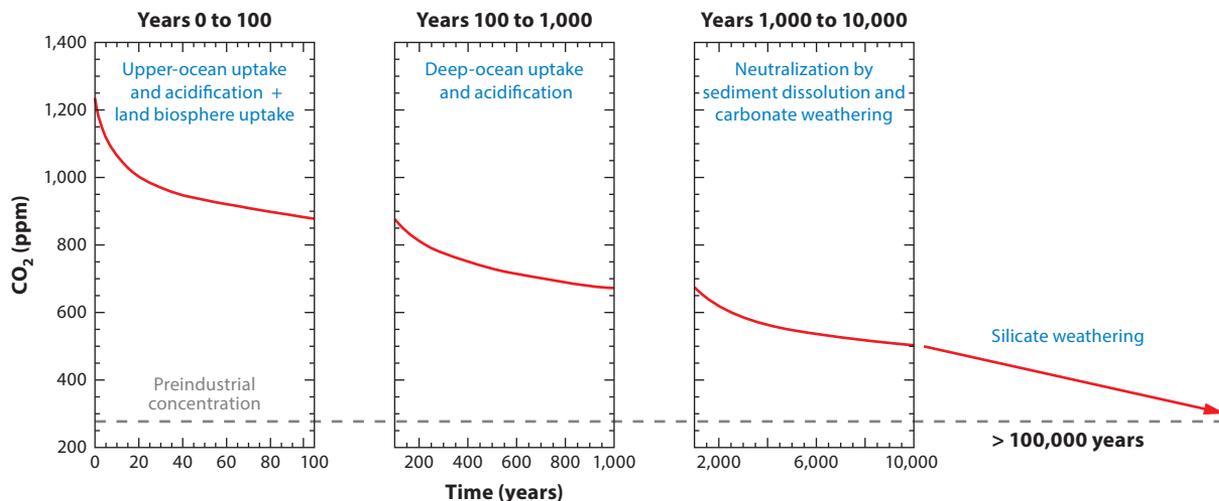


Figure 1

Decay of CO_2 concentration following instantaneous addition of 2,100 GtC to the atmosphere in the form of CO_2 . Based on calculations by Eby et al. (2009) (figure source: NRC 2011).

this nonlinearity has the effect that the proportion of emitted CO_2 that remains in the atmosphere at any given time t increases as the cumulative amount of CO_2 emitted up to time t increases. We refer to this as the air fraction nonlinearity, and it implies that on timescales of up to a few millennia, the atmospheric CO_2 concentration is concave upward as a function of the cumulative amount of CO_2 emitted.

The surface temperature of an earthlike planet can to a good approximation be inferred from the top-of-atmosphere radiation budget without detailed reference to what is going on in the rest of the atmosphere (see chapter 3 of Pierrehumbert 2010). In its simplest form, radiative forcing is the change in the top-of-atmosphere radiation budget caused by a specified change in an atmospheric constituent such as CO_2 , with temperature and all other meteorological inputs (including clouds and water vapor) held fixed. Changes in temperature structure, water vapor, and clouds are considered as part of the feedbacks determining the climate sensitivity, which translates radiative forcing into global mean surface temperature response (Soden & Held 2006). The stratosphere adjusts quickly to changes in greenhouse gas content, and the adjustment of stratospheric temperature alters the top-of-atmosphere budget and hence the surface temperature. This can be taken into account by computing top-of-atmosphere forcing after the stratosphere has been allowed to adjust. It has been found that the simple expedient of computing unadjusted radiative forcing at the top of the troposphere rather than at the top of the atmosphere yields very nearly the same result. When stated without further qualifiers, radiative forcing in this article always refers to the top-of-troposphere value. Even for greenhouse gases with geographically uniform concentrations the radiative forcing exhibits some geographical variation due to variations in temperature, clouds, and humidity, but because this variation is moderate in amplitude and smooth, and the atmosphere is quite good at mixing heat horizontally, global mean radiative forcing has proved to be a useful characterization of the drivers of climate change even for full general circulation models, as well as for the atmosphere itself. Once radiative forcing is known, the surface warming is determined by how much the atmosphere has to warm in order to bring the top-of-atmosphere radiation budget back into balance.

Air fraction: the proportion of emitted CO_2 that remains in the atmosphere

Radiative forcing: the change in top-of-atmosphere energy budget caused by changing an atmospheric constituent by a specified amount, while holding all other constituents and also temperature fixed

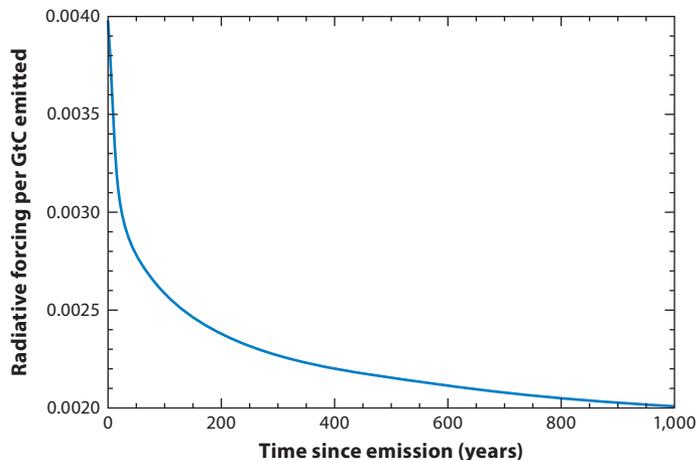


Figure 2

Time series of radiative forcing resulting from a pulse release of 1 GtC as CO₂, inferred from simulations in Eby et al. (2009). Because radiative forcing is nearly linear in the amount of carbon emitted, results for pulses of other magnitudes can be scaled linearly from these results.

Over the range of concentrations of interest in anthropogenic global warming, the radiative forcing (in W/m²) caused by changing atmospheric CO₂ concentration from a baseline value c_b to a value c is

$$\Delta F = 5.35 \ln \left(\frac{c}{c_b} \right). \quad (1)$$

This formula yields a radiative forcing of 3.7 W/m² per doubling of the CO₂ concentration. The coefficient in Equation 1 incorporates the effect of a fixed climatological distribution of clouds and water vapor, as well as the geographic variations of temperature. CO₂ radiative forcing is concave *downward* as a function of concentration. However, the air fraction nonlinearity makes CO₂ concentration concave *upward* as a function of cumulative emissions, and it is a somewhat fortuitous consequence of the nature of carbonate chemistry nonlinearities that these two nonlinearities very nearly cancel for cumulative emissions up to several thousand gigatonnes of carbon. As a result, the value of ΔF at the end of a given time period is linearly proportional to the cumulative emissions during that interval (Caldeira & Kasting 1993, Allen et al. 2009, Eby et al. 2009, Matthews et al. 2009, NRC 2011).

Figure 2 shows the proportionality factor between carbon emissions and the resulting radiative forcing, as a function of time since the emission took place. These results were derived from the pulse emission simulations in Eby et al. (2009) by doing least-squares linear fits between radiative forcing and the mass of carbon released. Details of the calculation are provided in the **Supplemental Material**.

The curve in **Figure 2** can be very accurately fit by the analytic form

$$b(t) = 0.004223(0.27218e^{-t/8.696} + 0.14621e^{-t/93.3} + 0.13639e^{-t/645.87} + 0.44522), \quad (2)$$

where t is the time in years since emission. CO₂ thus acts like a mixture of decadal-, centennial-, millennial-, and infinite-lifetime gases. Although CO₂ exhibits strong nonlinearities with regard to both radiative forcing per unit concentration and removal mechanisms, the linearity embodied by Equation 2 means that the net effects of CO₂ emissions can be regarded as due to a fictitious

equivalent “C-gas” with linear radiative forcing and decay curve given by Equation 2, independent of the mass of emissions. Given the pulse response function $b(t)$, the radiative forcing due to an arbitrary emission trajectory is

$$\Delta F(t) = \int_{-\infty}^t E(t')b(t-t')dt', \quad (3)$$

where $E(t')$ is the emission rate in GtC/year.

Throughout this article, we use a simple two-box energy balance model to translate time-dependent radiative forcing into temperature. The model consists of a shallow mixed layer with temperature perturbation T'_{mix} and heat capacity μ_{mix} coupled to a deep ocean with temperature T'_{deep} and heat capacity $\mu_{\text{deep}} \gg \mu_{\text{mix}}$. The mixed layer loses heat to space (in part via coupling to the atmosphere) at a rate proportional to its temperature. Relative to the unperturbed state, the mixed layer gains additional energy at a rate determined by the radiative forcing ΔF , which in the general case could include changes in the absorbed solar radiation as well as changes in outgoing infrared. The equations are

$$\begin{aligned} \mu_{\text{mix}} \frac{dT'_{\text{mix}}}{dt} &= -\lambda T'_{\text{mix}} - \gamma(T'_{\text{mix}} - T'_{\text{deep}}) + \Delta F(t) \\ \mu_{\text{deep}} \frac{dT'_{\text{deep}}}{dt} &= -\gamma(T'_{\text{deep}} - T'_{\text{mix}}). \end{aligned} \quad (4)$$

For constant radiative forcing ΔF , this model has the steady solution $T'_{\text{mix}} = T'_{\text{deep}} = \Delta F/\lambda$. Hence $1/\lambda$ gives the equilibrium climate sensitivity. The model relaxes to this equilibrium state on two timescales. On the short timescale (generally a matter of a few years), the mixed layer relaxes to a quasi-equilibrium but the deep ocean has not yet had time to warm up, so $T'_{\text{deep}} \approx 0$. The transient climate response during this stage is then $T'_{\text{mix}} = \Delta F/(\lambda + \gamma)$. If ΔF is reduced to zero some time after the deep ocean has warmed up to some nonzero value T'_{deep} , then on the short mixed-layer timescale T'_{mix} falls only to $T'_{\text{deep}} \gamma/(\lambda + \gamma)$ and subsequently relaxes to zero on the slow deep-ocean timescale. This is the recalcitrant warming due to heat burial in the deep ocean (Held et al. 2010). The standard parameters we use for this model are given in the caption to **Figure 3**.

Figure 3 shows the result of forcing the two-box model with the CO₂ time series resulting from the A2+ emissions scenario in Eby et al. (2009) that involves 1,700 GtC of emissions up to the year 2300 (at which time emissions cease). It illustrates another important feature of the climate response to CO₂ emissions: For climate sensitivity up to the top of the Intergovernmental Panel on Climate Change (IPCC) range, the warming remains essentially constant for the millennium following cessation of emissions. This property arises from the fact that the deep-ocean heat and carbon uptakes occur on similar timescales, so that the system warms toward equilibrium at the same rate CO₂ is drawn down toward equilibrium. This irreversibility of CO₂-induced warming was highlighted in Solomon et al. (2009). The flatness of the temperature response in the millennium following cessation of emissions (when climate sensitivity is not extremely high) has been confirmed in a number of intermediate-complexity coupled climate/carbon cycle models (Eby et al. 2009, Zickfeld et al. 2013), though it has not yet been extensively explored in fully coupled atmosphere-ocean general circulation models incorporating a carbon cycle. When the climate sensitivity becomes very high, however, instead of the temperature remaining flat, one encounters the phenomenon of committed warming, in which the warming grows substantially even after emissions have ceased. There are indications that in full atmosphere-ocean general circulation models long-term committed warming may occur at lower climate sensitivities as well (Frölicher et al. 2014).

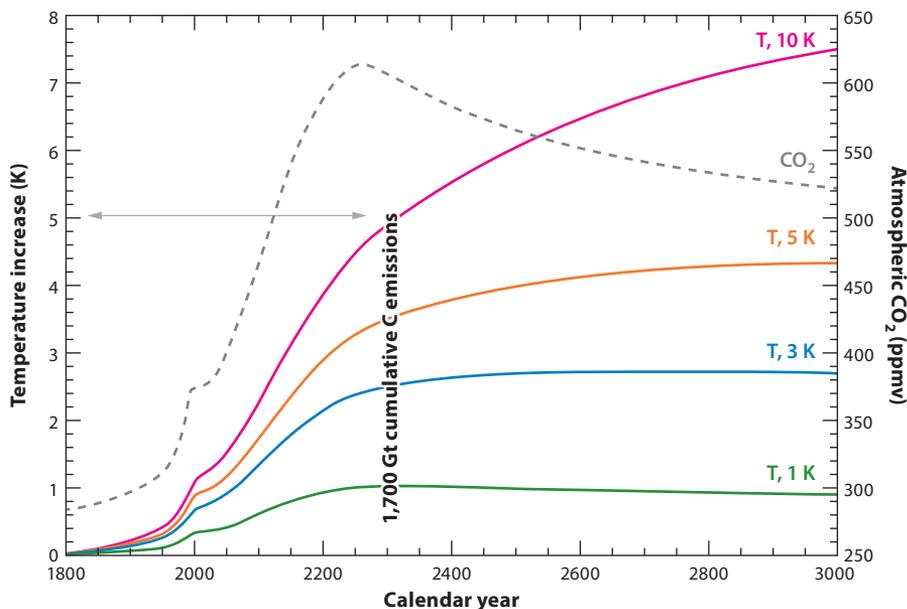


Figure 3

Warming resulting from driving the two-box climate model (Equation 4) with the radiative forcing from the A2+ emission scenario in Eby et al. (2009) that has total cumulative emissions of 1,700 GtC. The time series of CO₂ concentration is given by the dashed curve, and temperatures are given by the set of solid curves, labeled T. Results are shown for a range of values of equilibrium climate sensitivity, indicated by the numbers on the temperature curves as the warming that would result from a doubling of CO₂. Standard parameters are $\lambda = 1.2 \text{ W/m}^2/\text{K}$ (corresponding to 3°C equilibrium climate sensitivity), $\gamma = 1.2 \text{ W/m}^2/\text{K}$, $\mu_{\text{mix}} = 3.154 \times 10^8 \text{ J/m}^2/\text{K}$, and $\mu_{\text{deep}} = 6.307 \times 10^9 \text{ J/m}^2/\text{K}$. Climate sensitivity was varied by varying λ while holding other parameters fixed at their standard values.



Two additional properties of the climate response to cumulative carbon can be inferred from calculations of the type shown in **Figure 3**. First, the peak warming is linearly proportional to cumulative carbon up to the time emissions cease. This is a consequence of the linearity of radiative forcing in cumulative carbon. Second, the peak warming is nearly independent of the emissions trajectory, for fixed cumulative carbon. This arises from the fact that the decay in radiative forcing seen in **Figure 2** is approximately offset by the additional warming that occurs as the deep ocean approaches equilibrium. These properties have been confirmed in a wide variety of models (Allen et al. 2009, Eby et al. 2009, Matthews et al. 2009), and we have verified that the two-box model can reproduce them. The key properties of cumulative carbon are summarized in the sidebar, Cumulative Carbon Basics.

3. COMPARING THE EFFECTS OF WELL-MIXED GREENHOUSE GASES

3.1. Radiative Forcing and CO₂e

Radiative forcing provides the common currency through which the climate effect of various combinations of greenhouse gas concentrations can be compared. The well-mixed greenhouse gases affect the radiative forcing almost exclusively through their effect on the emission of infrared

CUMULATIVE CARBON BASICS

- Cumulative carbon is the net mass of carbon released to the atmosphere in the form of CO₂ by anthropogenic sources, up to a given time.
- There is no “sustainable” CO₂ emissions rate compatible with climate stabilization. Unless CO₂ emissions fall to essentially zero at some time, the climate will continue to warm indefinitely.
- The peak warming is linearly proportional to the cumulative carbon released up to the time CO₂ emissions cease and is nearly independent of the emissions trajectory.
- For IPCC median climate sensitivity, the proportionality constant is approximately 2°C per trillion tonnes carbon released.
- For climate sensitivities at the IPCC median or lower, the peak warming occurs at or slightly after the time emissions drop to essentially zero.
- The temperature at the time of peak warming is the temperature the climate remains stuck in for the millennium following the cessation of emissions.

radiation to space. The radiative forcing ΔF due to well-mixed greenhouse gases is a nonlinear function of the concentration increments δc_j of the set of well-mixed greenhouse gases present in the atmosphere. The concentrations are assumed uniform throughout the atmosphere, and the increments are taken relative to some baseline set of concentrations, typically representing preindustrial values. One can define the linearized radiative efficiency—often simply called radiative efficiency—of gas j as

$$a_j \equiv \left. \frac{\partial \Delta F}{\partial c_j} \right|_{(c_0, c_1, \dots, c_n)}, \quad (5)$$

where (c_0, c_1, \dots, c_n) are the concentrations in the unperturbed atmosphere. With this definition, the radiative forcing is $\Delta F = \sum a_j \delta c_j$ if the concentration changes are sufficiently small. The radiative efficiencies depend on the set of baseline concentrations used in the linearization. Based on background values of 378 ppm for CO₂ and 1.75 ppm for CH₄ prevailing circa 2005, the radiative efficiency of CO₂ is 1.4×10^{-5} W/m²/ppb while that of CH₄ is 3.7×10^{-4} W/m²/ppb, or a factor of 26 greater.

Gases present in low concentrations tend to have higher radiative efficiency than gases like CO₂ that start off with relatively high concentration (**Supplemental Material**). CH₄ has a much higher radiative efficiency than CO₂ simply because its concentration is low, and not because its molecular structure makes it an intrinsically better greenhouse gas (Pierrehumbert 2010, section 4.5.4). The use of linearized radiative efficiency always overestimates the radiative forcing when concentration changes are large enough. Among the well-mixed greenhouse gases, only CO₂, CH₄, and N₂O exhibit significant nonlinearity over the range of anthropogenic concentration changes. (See **Supplemental Material**.)

The aggregate radiative forcing due to a mix of greenhouse gases is sometimes expressed as CO₂-equivalent or CO₂e. This is defined as the concentration of CO₂ that would yield the same radiative forcing as the full set of gases (including the CO₂), if CO₂ were the only source of radiative forcing (**Supplemental Material**). Because of the nonlinearity of CO₂ radiative forcing, CO₂e is not an additive quantity—it cannot generally be decomposed into contributions from the individual gases going into the mix. Note also that CO₂e defined in this way is a concentration-equivalent metric, characterizing the effects of the concentrations prevailing at a given time. It must not be confused with the emission-equivalent metrics, which regrettably go by the same

 **Supplemental Material**

Radiative efficiency: the change in radiative forcing per unit change in concentration of a given atmospheric constituent. Usually measured in W/m²/ppb

Indirect radiative efficiency: radiative efficiency taking account the radiative forcing due to changes in atmospheric chemistry caused by an emitted substance

Efficacy (or radiative efficacy): the global mean surface temperature change caused by 1 W/m^2 of radiative forcing due to an emitted substance, measured relative to that of CO_2

name but attempt to characterize the climate change caused by the emission of a given set of masses of greenhouse gases, as discussed in Section 3.4.

The radiative efficiency defined above is known as the direct radiative efficiency. If there are fast chemical feedbacks that alter the atmospheric composition in response to introduction of a greenhouse gas, then including these in the calculation of radiative forcing defines the indirect radiative efficiency. Addition of methane to the atmosphere increases stratospheric water vapor because methane breaks down into water there. Also, by depleting oxidizing agents in the troposphere, methane increases tropospheric ozone. Both effects add to the radiative forcing of methane alone and increase its radiative efficiency by about 35% (Hansen et al. 2005), though the estimate is model dependent and subject to considerable uncertainty. Climate system feedbacks, such as changes in tropospheric water vapor or clouds, are never included in radiative efficiency, because they respond to the changing state of the climate as a whole and not to any one climate forcing agent in isolation.

Shindell et al. (2009) argues for further increasing the indirect radiative efficiency of methane, on the grounds that methane inhibits the formation of sulfate aerosol. To include such a feedback in policy analysis, however, would be in effect to consider it a benefit that reducing methane makes sulfate pollution worse. That would make as little sense as paying power producers to use high-sulfur coal because of the cooling effect of the resulting sulfate aerosols, at the same time that other policy vehicles are attempting to reduce sulfate pollution in order to reduce acid rain, or to eliminate coal burning (and associated sulfate pollution) entirely for the sake of CO_2 mitigation.

The well-mixed greenhouse gases differ somewhat in the geographical patterns of the radiative forcing they cause, because of differences in the way the radiative forcing is affected by the temperature, water vapor, and cloud distributions. This can lead to differences in the amount of warming caused by a given global mean radiative forcing. The factor characterizing the amount of global mean surface warming caused by a fixed global mean radiative forcing, relative to that caused by CO_2 , is called the efficacy (Hansen et al. 2005). The efficacy for well-mixed greenhouse gases is quite close to unity; for example, it has a value of 1.1 for methane. Efficacy can be computed for any form of climate forcing agent, including sulfate aerosols and black carbon. Sometimes the efficacy is computed on the basis of indirect radiative efficiency, in which case it should be distinguished by the term indirect efficacy. The indirect efficacy of methane is estimated to be 1.45 (Hansen et al. 2005).

3.2. A Taxonomy of Well-Mixed Greenhouse Gases

Table 1 presents a taxonomy of well-mixed greenhouse gases, organized mainly according to lifetime. The Montreal gases—ozone-destroying gases controlled by the Montreal Protocol—are a special case. They currently have appreciable radiative forcing, but their emissions are already scheduled to go to zero, and so they do not present opportunities for further abatement. The well-mixed greenhouse gases are further distinguished by differences in their radiative efficiencies. The radiative efficiency mostly follows the typical concentration of the gas. CO_2 is present at a level of hundreds of parts per million and has the lowest radiative efficiency, methane and N_2O are present at hundreds of parts per billion and have intermediate radiative efficiencies, and all the gases with radiative efficiencies of 0.1 or more have concentrations of on the order of a hundred parts per trillion or less. To provide an idea of the magnitude of the climate impact of the various gases, and the potential for abatement, we also show the radiative forcing in 2010, based on measured atmospheric concentrations, and two projections of radiative forcing in 2100 based on the highest-end representative concentration pathway (RCP) emissions scenarios, RCP6 and RCP8.5 (van Vuuren et al. 2011), used in the IPCC Fifth Assessment Report (AR5). These scenarios provide a general indication of the plausible potential for growth but do not completely explore what

Table 1 Taxonomy of well-mixed greenhouse gases^a

Category	Gas(es)	Lifetime (years)	Radiative efficiency (W/m ² /ppb)	ΔF (W/m ²) (2010) ^b	ΔF (W/m ²) (2100) ^b	Remarks
CO ₂	CO ₂	NA	1.4 × 10 ⁻⁵	1.76	4.67–6.46	Cannot be characterized by a single lifetime, but significant effects persist for millennia
Millennial trace gases	SF ₆	3,200	0.52	0.0036	0.021–0.026	
	PFC-14 CF ₄	50,000	0.1	0.0083	0.015–0.022	
	PFC-116 C ₂ F ₆	10,000	0.26	0.0010	0.0035–0.0037	
	C ₆ F ₁₄	3,200	0.49	3.5 × 10 ⁻⁵	0.0001–0.0001	
	Other perfluorinated	740–4,100	0.26–0.57	≈0?	≈0?	
Centennial gases	N ₂ O	114	3.0 × 10 ⁻³	0.168	0.414–0.493	Mostly agricultural; limited abatement opportunities
	HFC-23	270	0.19	0.004	0.004	Mainly HCFC by-product
	HFC-236fa	240	0.28	≈0?	≈0?	Fire suppressant
	HFE-125	136	0.44	≈0?	≈0?	A fluorinated ether
Decadal gases	CH ₄	12	3.7 × 10 ⁻⁴	0.483	0.434–1.08	Methane
	Most HFCs	1.4–52	0.09–0.4	0.016	0.043–0.178	CFC replacements
	Most HFEs	0.33–26	0.25–1.37	≈0?	≈0?	Fluorinated ethers
Montreal gases ^c	CFCs	45–1,700	0.18–0.32	—	—	Phase-out scheduled
	HCFCs	1.3–17.9	0.14–0.32	—	—	Phase-out scheduled
	Others	0.7–65	0.01–0.32	—	—	Phase-out scheduled
	All	0.7–65	0.01–0.32	0.317	0.080	Phase-out scheduled

Abbreviations: CFC, chlorofluorocarbon; HCFC, hydrochlorofluorocarbon; NA, not applicable; PFC, perfluorocarbon.

^aLifetimes and radiative efficiency are taken from Forster et al. (2007); the radiative forcing in 2010 and range of radiative forcing in 2100 are taken from the representative concentration pathway (RCP) database (van Vuuren et al. 2011), with range based on RCP6 to RCP8.5.

^bThe symbol ≈0? indicates that precise radiative forcing estimates could not be found in the literature, but are likely to be negligible.

^cRadiative forcings are not broken out by component for Montreal gases, but are only given in the aggregate.

might happen in the total absence of climate protection actions, though RCP8.5 is a scenario with largely unrestrained emissions and may be taken as a rough approximation to that case.

CO₂ is in a class by itself, for the reasons discussed in Section 2. The perfluorocarbons (PFCs) and SF₆ have millennial lifetimes and high radiative efficiency but are present in such tiny quantities that, even allowing for emissions growth, they contribute little radiative forcing in comparison with CO₂. Though reduction in millennial trace gas emissions is significant at the margins, it is our emissions of CO₂ in the coming century that will determine the character of the Anthropocene. On centennial timescales, N₂O is by far the dominant player, but its primary sources are intrinsically linked to the essential role of nitrogen in agriculture, so opportunities for abatement are very limited (Davidson 2012). The HFCs are ozone-safe replacements for chlorofluorocarbons (CFCs); a few are centennial gases, but most are decadal. HFC-23, which is the most abundant centennial HFC, should properly be considered a Montreal gas, because it is primarily a by-product of hydrochlorofluorocarbon (HCFC) production, and emissions are likely to cease by 2020. Methane is by far the dominant decadal gas and is the only one that presents opportunities for significant abatement of current emissions. Abatement of the entire present anthropogenic methane emissions

Anthropocene: the period of geologic time during which human activities play a major role in altering the climate and biogeochemistry of Earth

Short-lived

greenhouse gas: for the purposes of this review, a greenhouse gas having an atmospheric lifetime of up to a few decades

would remove 0.483 W/m^2 of direct radiative forcing, and under a high-emissions scenario the radiative forcing potentially available for abatement grows to just over 1 W/m^2 . However, nearly 25% of current methane emissions and 37% of the year-2100 high-end emissions are attributable to extraction and use of fossil fuels (Riahi et al. 2007, van Vuuren et al. 2011) and would be subject to abatement as an automatic cobenefit of reducing CO_2 emissions. The main opportunities for abatement unrelated to fossil fuel usage lie in agricultural sources, which make up 36% of the high-end total, and waste handling (including landfills), which makes up 24% of the high-end total.

The decadal HFCs currently contribute negligible radiative forcing, but increased usage, primarily through expansion of refrigeration and air conditioning in the developing world, could increase the radiative forcing to significant levels in the future. In the high-emissions RCP8.5 scenario, the radiative forcing grows to 0.178 W/m^2 by 2100. Assuming high economic growth rates in the developing world and completely unrestrained HFC emissions, Velders et al. (2009) estimate that the radiative forcing could be up to 0.3 W/m^2 by 2050.

The term short-lived greenhouse gas is sometimes used to refer to gases like ozone or water vapor that have a residence time in the atmosphere of a month or less and therefore do not become well mixed. For the purposes of this review, however, we expand the definition of short-lived to encompass gases like methane or HFCs, which have lifetimes of up to a few decades. The main distinction in this review is between such gases and gases like CO_2 or SF_6 , which have climate effects of millennial or longer duration. Centennial gases are in a gray area between the two major categories, but because no centennial gases present major opportunities for new abatement, the problem is not one that need concern us greatly.

3.3. Effects of Greenhouse Gas Lifetime

Climate protection protocols are implemented by setting limits on *emissions*, not concentrations. To compare the climate impact of emissions it is necessary to compute the corresponding effect on concentration, and this requires consideration of the sink of the gas, which can be characterized by atmospheric lifetime for gases whose sinks lead to a simple exponential decay. The chief difficulty in comparing the climate impact of the various well-mixed greenhouse gases arises from the need to take into account the differences in their atmospheric lifetimes, which span the range summarized in **Table 1**. CO_2 cannot be characterized by a single lifetime, but if one wants to situate CO_2 in the lifetime hierarchy, it can be regarded as a mix of decadal-, centennial-, millennial-, and infinite-lifetime gases with corresponding radiative forcings being linear in emissions; 58% of emissions go into the millennial- to infinite-lifetime component (Equation 2). A more precise comparison of CO_2 with other gases requires the use of a suitable carbon cycle model, as is done in Sections 3.4.1 and 5.

Let M be the atmospheric inventory of a substance, typically expressed as total mass of the substance in the atmosphere. For a substance that can be characterized by exponential decay with lifetime τ , $M(t)$ is related to emissions rate $E(t)$ by the differential equation

$$\frac{dM}{dt} = -\frac{1}{\tau}M + E(t), \quad (6)$$

where it is presumed that E and M are measured in consistent units. The radiative forcing depends on the atmospheric concentration c , which for Earth's atmosphere is related to the inventory by $c = (5,500.34/m)M$, where m is the molecular weight of the gas, M is the atmospheric inventory in units of Gt, and c is expressed in ppb. Sometimes the emission rate and inventory are stated in terms of the mass in Gt of one key element in the molecule, rather than the mass of the gas as a whole, in which case one simply uses the atomic weight of the element for m instead of the

molecular weight of the gas. This is most commonly done when keeping track of the carbon in CO₂. Thus, adding 1 GtC in the form of CO₂ to the atmosphere increases the CO₂ concentration by (5,500.34/12) or 458 ppb, but adding 1 Gt of CO₂ increases the concentration by (5,500.34/44) or 125 ppb. When linearization is valid, one multiplies c by radiative efficiency a in the appropriate units (W/m²/ppb) in order to obtain the radiative forcing ΔF . In the remainder of the discussion in this section, we assume the linear approximation to radiative forcing, so as to focus on the trade-off between lifetime and radiative efficiency. The radiative efficiency in units of W/m² per Gt added to the atmospheric inventory is then $a' = (5,500.34/m)a$.

Suppose now that $E(t)$ varies with a characteristic timescale τ_E , which we can think of as being a century or two in the case of most anthropogenic emissions. Then, we can distinguish two limiting cases: the short-lived case with $\tau \ll \tau_E$ and the long-lived case with $\tau \gg \tau_E$. In the short-lived case, $M(t) \approx E(t)\tau$, so $\Delta F \approx (a'\tau)E(t)$. Thus, in the short-lived case, radiative forcing responds to the *current emissions rate* and goes up and down as the emission rate is increased or decreased. In this sense, the climate forcing due to short-lived gases is *reversible*. One can eliminate it at any time by reducing the emission rate to zero, and get back to zero radiative forcing. This contrasts with the long-lived case, for which

$$\Delta F \approx a' \int_0^t E(t) dt. \quad (7)$$

The radiative forcing responds to the *cumulative emissions*. Bringing the emission rate down to zero does not bring the radiative forcing back down to zero. CO₂ is a good example of a long-lived gas in this sense, whereas methane is a good example of a short-lived gas, with N₂O lying between the two in behavior.

To underscore the contrasting response of climate to short- versus long-lived greenhouse gases, in **Figure 4** we show the result of driving the climate model in Equation 4 with an emissions trajectory $E = E_m \exp(-(t - t_m)^2/100^2)$, which rises to a maximum E_m at time t_m and falls symmetrically to zero with a timescale of 100 years. The results depend on the maximum emissions rate and radiative efficiency only through the product $a'E_m$, which is held fixed in the simulations as the lifetime of the gas is varied.

The temperature response to short-lived gases shows the expected nearly symmetric rise and fall, with a slight lag between peak warming and peak emissions, which increases as the lifetime of the gas increases. For gases with a lifetime of 20 years or less, the peak warming increases very nearly in proportion to lifetime τ , because (for fixed radiative efficiency) the concentration yielded by a given emissions rate increases in proportion to τ . Because the timescales involved are short compared with the deep-ocean equilibration time, the peak radiative forcing is translated into temperature change using the transient climate sensitivity, which is 0.42°C per W/m² in this simulation; the predicted warming is then $0.42a'E_m\tau = 0.0042\tau$ for the short-lived gases, which is consistent with the results in the figure. The peak warming predicted by this scaling remains approximately correct even for the gas with a 100-year lifetime, though the position of the peak is lagged fully 100 years past the emissions peak in that case. One can see a tail of warming that decays more slowly than the greenhouse gas itself. This is the recalcitrant warming due to heat storage in the deep ocean (Solomon et al. 2010, Winton et al. 2010). It is present for the shorter-lived gases as well, though less evident because of the scaling of the graph.

The long-lived gases exhibit the expected irreversibility of warming, which persists long after emissions cease. For the case with 1,000-year lifetime, the peak warming occurs at about the time emissions approach zero and declines only modestly over the subsequent millennium. For still longer-lived gases, the temperature exhibits committed warming—a continued increase in temperature even after emissions cease, as the deep ocean equilibrates with a nearly unchanging

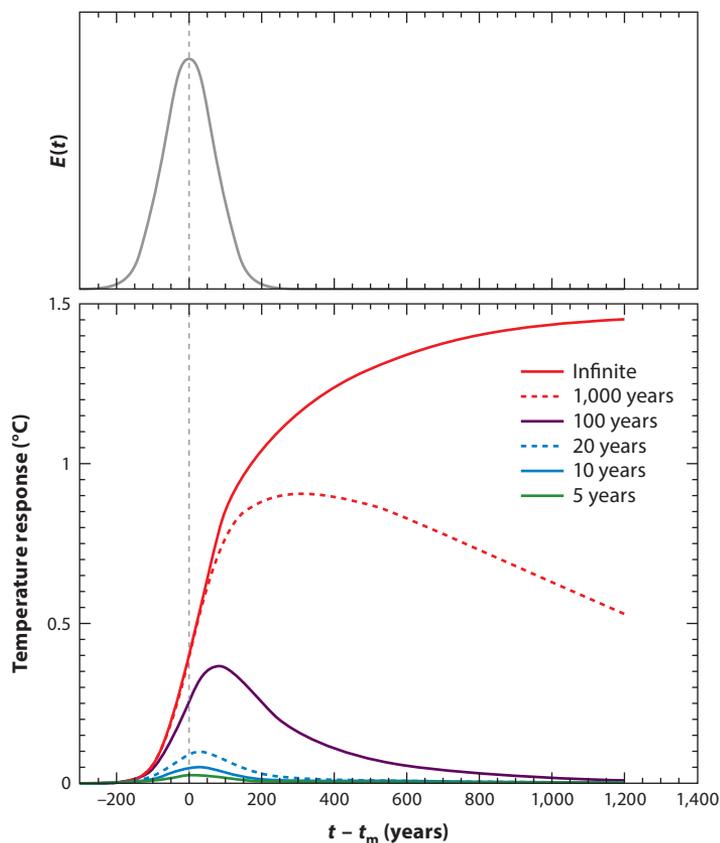


Figure 4

Temperature response (*lower panel*) of the two-box climate model to an emission profile that rises to a maximum value E_m and falls symmetrically to zero, as illustrated in the upper panel. Results are shown for various values of the greenhouse gas lifetime, with $a'E_m$ held constant at $0.01 \text{ (W/m}^2\text{)/year}$, where a' is the radiative efficiency in mass units.

radiative forcing. In contrast to the short-lived gases, for which the peak warming is largely determined by the transient climate sensitivity, the long-lived gases are persistent enough that the resulting warming is governed by the larger equilibrium sensitivity.

The upshot is that the climate impact of short-lived gases can be well characterized by weighting the emission rate by the factor $a'\tau$, which captures the trade-off between lifetime and radiative efficiency. This serves as a precise basis for comparing short-lived gases *with each other*. For policy guidance, this statistic must be augmented with the understanding that the climate response to short-lived gases depends primarily on *current* emissions, with little memory of the past. In contrast, the climate response to long-lived gases is characterized by their cumulative emissions over all past times, weighted by a' , and the resulting climate impact is essentially irreversible. Because of the fundamentally different nature of the climate response to long- versus short-lived gases, there is no way to express emissions of short-lived gases in terms of an equivalent in emissions of long-lived gases without seriously misrepresenting some aspect of the climate response.

The essential irreversibility of the warming caused by long-lived greenhouse gases has often led to the misconception that mitigation of short-lived gases is somehow intrinsically preferable,

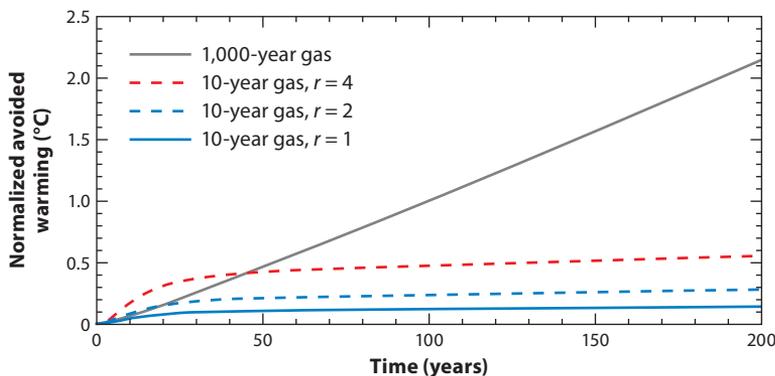


Figure 5

Avoided warming due to reduction of emission rate of a greenhouse gas by an amount ΔE . The graph shows the comparison between results for a gas with 1,000-year lifetime and three different values of ΔE for a gas with a 10-year lifetime. The emission reduction for the short-lived gas relative to the long-lived gas is characterized by the parameter r , which is the ratio of $a' \Delta E$ for the short-lived gas to that for the long-lived gas, where a' is the radiative efficiency of the respective gas expressed in mass units. The value of ΔE for the long-lived gas is chosen to yield an avoided warming of 1°C after 100 years.

in that reducing emissions of short-lived gases leads to a nearly immediate reduction in radiative forcing. **Figure 5** demonstrates the fallacy of this point of view. Based on Equation 4, we show the warming that would be avoided by reducing annual emissions rates by a time-independent amount ΔE relative to a base case. Results are shown for a methane-like gas with a lifetime of 10 years and a CO₂-like gas with a lifetime of 1,000 years, though over the time span shown in the graph the difference between a 1,000-year and infinite lifetime is immaterial. The results depend on ΔE only through the product $a' \Delta E$. The value of $a' \Delta E$ for the long-lived gas is normalized to a value that yields 1°C avoided warming after 100 years, and the comparative avoided warming for the short-lived gas is determined by the ratio r of $a' \Delta E$ for the short-lived gas to that for the long-lived gas. The essential difference in character between the decadal case and the millennial case is that the avoided warming from the decadal gas quickly saturates at a maximum value, whereas the avoided warming from the long-lived case grows in time without bound. There is some growth in the avoided warming in the decadal case, owing to warming of the deep ocean, but the effect is slight. When $r = 1$, the avoided warming for the millennial case is always greater than that for the decadal case, so given the choice it is better to reduce long-lived than short-lived emissions regardless of what time frame one may consider most important. For $r > 1$, there is a brief period during which the avoided warming for the decadal case exceeds that for the long-lived case, but this temporary advantage comes from having abated a larger amount of emissions (weighted by radiative efficiency) and is not an advantage conferred by the short-lived nature of the gas. In the long run—which is generally not very long at all—abatement of the long-lived gas always yields more avoided warming than abatement of the decadal gas. For $r = 4$, abatement of the decadal gas yields more avoided warming out to 45 years. In that case, abatement of the short-lived gas is the preferable alternative only if one's goal is to minimize warming 45 years or less in the future, and one cares nothing at all what happens at later dates—even if choosing the short-lived option leads to a considerable increase in peak warming, as it will if the peak warming occurs more than 45 years out. To relate these numbers to the methane versus CO₂ decision, one may use Equation 2 to infer that the value of $a' \Delta E$ for the millennial and longer contribution of abatement of 10 GtC/year of CO₂ is 0.0246,

corresponding to the long-term effect of abating all of the present CO₂ emissions. Abating all of the present methane emissions yields $a' \Delta E = 0.054$, allowing for an efficacy and feedback factor of 1.5. This yields $r = 2.2$. Taking a more plausible case wherein one abates half of the current methane emissions but only a quarter of the current CO₂ emissions raises r to 4.4, not very different from the $r = 4$ case shown in the graph. Even this comparison is excessively favorable to the value of methane abatement, because it neglects the part of methane abatement that is an automatic cobenefit of CO₂ abatement and the radiative forcing due to the decadal and centennial contributions of CO₂.

The intuitive interpretation of the preceding is that the warming due to millennial gases such as CO₂ is irreversible, but avoidable. The advantage of CO₂ abatement over abatement of short-lived climate forcing grows with time.

Figure 5 does not imply that the optimal strategy is to abate as much methane as possible until the crossover point is reached and then switch emphasis to CO₂ abatement. To get the long-term advantage of CO₂ abatement, one must start early, to give time for its advantage

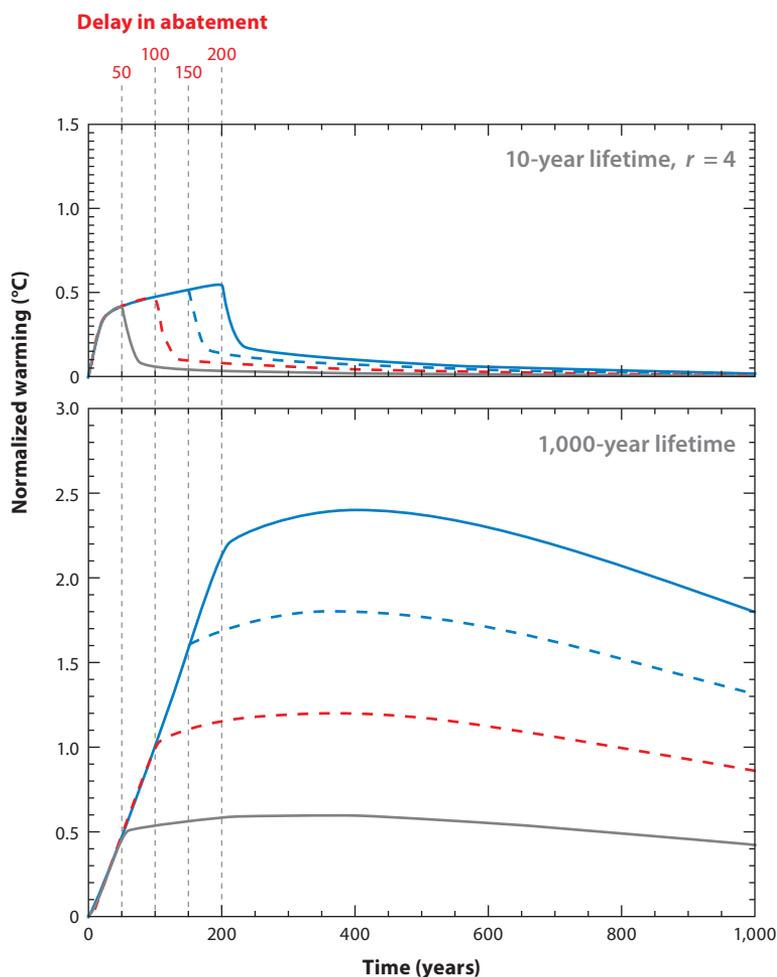


Figure 6

Consequences of delay in mitigation of a decadal- versus millennial-lifetime gas. Normalization is as for **Figure 5**. Calculations are done with $r = 4$.

to grow. This point is illustrated in **Figure 6**, which illustrates the consequences of delay in mitigation of a short-lived versus a long-lived gas, for the case $r = 4$. There is little harm in delaying the mitigation of the decadal gas, because almost all of the advantages of mitigation will be captured within a decade of the time mitigation occurs. Hence, there is little to be gained from early mitigation of the short-lived gas. In contrast, any delay in mitigation of the long-lived gas ratchets up the warming irreversibly. But to get the full benefit of CO₂ abatement that accrues by year 100, one must begin in year zero. The situation is rather like saving money for one's retirement—the earlier one begins saving, the more one's savings grow by the time of retirement, so the earlier one starts, the easier it is to achieve the goal of a prosperous retirement. Methane mitigation is like trying to stockpile bananas to eat during retirement. Given the short lifetime of bananas, it makes little sense to begin saving them until your retirement date is quite near.

3.4. Emission Metrics

An emission metric is a function that takes as input a set of masses of various greenhouse gases and produces as output a single number intended to represent the harm (via climate change) caused by the emissions of those gases. To be useful, an emission metric needs to be linear; otherwise, the value of the metric cannot be decomposed into contributions from individual emitting entities such as nations, or even from emissions occurring at various times. Thus, a metric generally takes the form of a weighted sum of the masses of the emitted gases. The weights, however, can be functions of time (as in Section 3.4.2). Because radiative forcing is linear in cumulative CO₂ emissions, and also linear in emissions of trace gases such as HFCs and PFCs, linearity does not pose a problem for these gases if climate-related damages are linear in radiative forcing. The nonlinearity of radiative forcing in methane and N₂O emissions does pose a slight problem, but this can be addressed by making the weights of these gases dependent on the concentrations prevailing in a given year. The resulting metric then becomes quasi-linear rather than truly linear, but this is not too problematic so long as the concentrations do not change rapidly within a given year.

Emission metrics are intended to simplify the formulation of climate-control vehicles such as emissions limitation treaties, taxes, or emissions trading schemes; a good metric would allow one to avoid having to write specific provisions dealing with each individual gas to be controlled. Metrics are also used to simplify analysis of climate policies, though we argue in Section 3.4.3 that the use of metrics can and should be avoided for such purposes.

In light of the results of Section 3.3, emission metrics that aggregate short-lived and long-lived gases seek to do the impossible, because the two kinds of gases have fundamentally different consequences for climate. We argue that the enterprise of seeking a metric to aggregate long-lived and short-lived gases is fundamentally misconceived and should be abandoned in favor of sets of metrics that aggregate gases in similar lifetime classes.

The most widely adopted form of climate protection goal is expressed as a temperature threshold not to be exceeded, stated either at a fixed time (e.g., up to the year 2100) or as a peak warming regardless of when the peak occurs (climate stabilization). Therefore, in evaluating metrics, the extent to which the metric provides a guide to the achievement of this form of goal deserves particular attention. More generally, we need to look at the extent to which sets of emissions with equal metric value yield different climate futures.

3.4.1. Global warming potential. The only emission metric in widespread use at present is the global warming potential (*GWP*), which is a crude attempt to characterize the aggregate effect of greenhouse gases with widely disparate lifetimes. The definition of *GWP* begins with the premise that the damages caused by emission of a greenhouse gas are proportional to the

associated radiative forcing, averaged over a prescribed time interval. It then layers on a number of additional assumptions:

- The concentration change $\delta c(t)$ used to compute radiative forcing is that caused by a pulse emission of the gas at time t_0 .
- The concentration change is linear in the mass of the gas emitted.
- The radiative forcing is linear in the concentration change.
- The radiative forcing is averaged over an interval of fixed, specified duration b , from the time of emissions t_0 to time $t_0 + b$.

The last of these is immediately problematic, because it means that GWP is completely insensitive to any damages caused by climate change that occurs more than a time b following the emission, no matter how catastrophic they may be.

The GWP (relative to CO_2) for gas A over time horizon b is then defined as (Forster et al. 2007)

$$GWP_b = \frac{\overline{\Delta F}^{(b)}(\text{A})}{\overline{\Delta F}^{(b)}(\text{CO}_2)}, \quad (8)$$

where $\overline{\Delta F}^{(b)}$ is the radiative forcing averaged over time horizon b , and the concentration trajectory used in computing the radiative forcing is that resulting from a pulse emission of a mass ΔM of each gas at the beginning of the interval. Because of the linearity assumptions, the value chosen for ΔM is immaterial. GWP is conventionally stated relative to the mass of CO_2 , rather than carbon, emitted. The GWP_b emission metric is the sum of emissions of all gases, weighted by their respective GWP_b .

GWP_b for a short-lived gas relative to CO_2 or any other long-lived gas is a sensitive function of the time horizon b . When b becomes larger than the lifetime τ of the short-lived gas, the numerator of Equation 8 goes to zero like $a'_A \tau \Delta M / b$, whereas the denominator remains approximately fixed at $a'_{\text{CO}_2} \Delta M$, whence the GWP approaches zero as b is made larger.

It can be shown that for a gas with lifetime much shorter than b , $\overline{\Delta F}^{(b)}$ is independent of the emissions trajectory, so the pulse emission assumption used in defining GWP is not problematic in that case. However, for a longer-lived gas, $\overline{\Delta F}^{(b)} \approx a' \int_0^b (1 - \frac{t}{b}) E(t) dt$, which weights early emissions more than late emissions (**Supplemental Material**). The dependence of GWP on emissions trajectory can lead to serious inaccuracies when comparing short-lived to long-lived greenhouse gases, as was noted by Smith & Wigley (2000b).

The mean temperature over time interval b is proportional to $\overline{\Delta F}^{(b)}$ (**Supplemental Material**). In consequence, even apart from the inaccuracies in the estimate of $\overline{\Delta F}^{(b)}$, GWP_b is unsuitable for implementation of temperature threshold or peak temperature targets; it characterizes the mean warming over the interval out to time b , but is insensitive to whether the warming occurs at the beginning or end of the interval, or at a late enough time that the warming affects the peak temperature. Any climate-control protocol based on GWP_b will unduly reward early mitigation of short-lived gases, given that such mitigation contributes nothing to achieving the climate protection goal.

To illustrate the preceding points in a more concrete fashion, we now present three calculations demonstrating that emissions pathways that are identical from the standpoint of GWP can yield very different climate outcomes. These examples are carried out for the case of GWP_{100} .

In **Figure 7** we show the warming caused by two different greenhouse gases, each of which is emitted at a constant rate for 200 years, followed by instantaneous cessation of emissions; one gas has a lifetime of 10 years, and the other a lifetime of 1,000 years, and the emissions rates are chosen such that the two gases have equal net emissions weighted by GWP_{100} . Emission of the short-lived

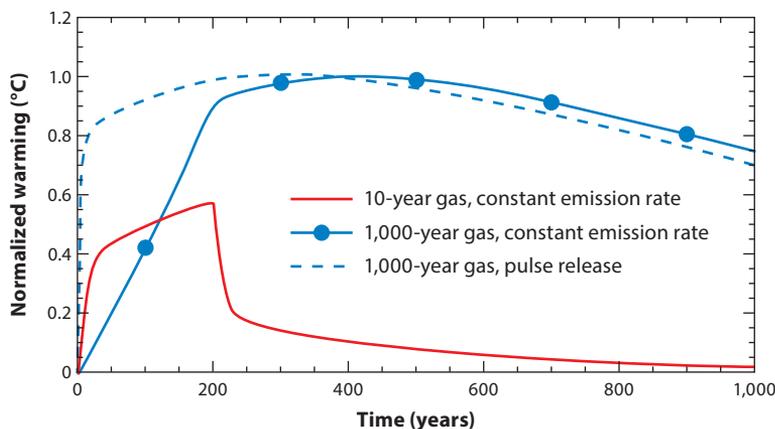


Figure 7

Effect of lifetime on warming caused by emissions that have equal GWP_{100} . Solid lines show results for constant-rate emissions of a 10-year and 1,000-year gas, which begin at year 0 and terminate at year 200. The dashed line shows results for an instantaneous pulse emission in the 1,000-year case. Temperatures are normalized such that the constant-emissions case for the 1,000-year gas yields a 1°C peak warming.

gas initially yields somewhat more warming than does emission of the long-lived gas, but by the time emissions cease, the long-lived gas yields nearly twice the warming yielded by the short-lived gas. Moreover, the warming caused by the long-lived gas persists long after the emissions cease, whereas that from the short-lived gas mostly dissipates within a few decades. These two climate futures are very different, and it is hard to think of any criterion by which the climate future with larger and irreversible warming is the preferable alternative.

In **Figure 7** we also show what happens if the emission of the long-lived gas occurs in a pulse at the beginning of the period, instead of being spread out over 200 years. In that case, the warming from the long-lived gas exceeds that from the short-lived gas at all times, even though the pulse emission has the same GWP as the sustained emission. This illustrates the sensitivity of climate response to the emissions trajectory for long-lived gases and again shows that emissions with identical GWP can yield substantially different climate futures.

Next, in **Figure 8**, we consider the climate response to two different emissions trajectories of a gas with a 10-year lifetime. Scenario E_1 has constant emissions at a low rate beginning in year 0 and terminating after 50 years, while the emissions in E_2 are at a higher rate but only begin in year 80 and terminate after 20 years, yielding identical GWP -weighted emissions to E_1 . This example is to be thought of in the context of a climate protection target expressed as a temperature threshold to be met at year 100. Scenario E_2 yields nearly twice the peak warming of scenario E_1 , and moreover contributes this larger warming to the temperature near year 100, and therefore greatly affects the target. Any protocol based on GWP , however, would assign equal merit to abating the emissions in scenario E_1 , even though that abatement would contribute very little to the achievement of the target. This shortcoming of GWP and similar emission-equivalent metrics was noted by Manne & Richels (2001).

Figure 9 shows how the short-lived versus long-lived contrast plays out in a more realistic setting. The CO_2 time series used in these calculations are drawn from the extended A2 family described in Eby et al. (2009) and result from driving a carbon cycle model with historical emissions up to the year 2000, followed by an emission rate that peaks in the year 2100 and thereafter declines linearly to zero by the year 2300; the members of the family are labeled according to the cumulative

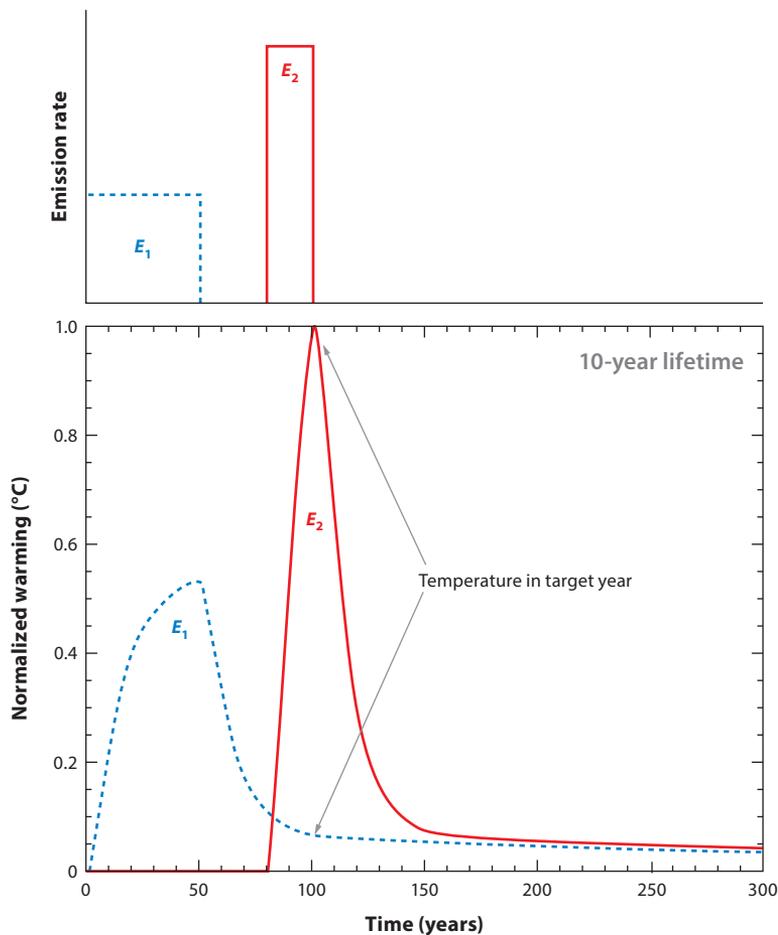


Figure 8

Warming due to emission of a gas with 10-year lifetime, following the two emission trajectories in the upper panel. The two cases have equal GWP_{100} . Temperatures are normalized such that case E_2 yields 1°C peak warming.

carbon emitted between 2000 and 2300. The corresponding radiative forcing is computed from Equation 1. These calculations are essentially the same as those discussed in Solomon et al. (2012), except that these are done with a climate sensitivity of 3°C per doubling. The solid red curve shows the temperature obtained by driving Equation 4 with the CO_2 from the 1,280-GtC scenario alone. The dashed blue curve shows the result of forcing the model with the CO_2 from the lower-emission 640-GtC scenario, but with additional forcing due to methane emitted at a constant rate beginning in the year 2000 and ending in 2300. The methane emission rate is chosen such that the net methane emissions weighted by GWP_{100} are equal to the 640-GtC difference in the CO_2 emissions between the two cases. Thus, the two cases shown in the figure have identical GWP_{100} -weighted emissions. The GWP was taken from table 2.14 of Forster et al. (2007) and includes the effect of ozone and stratospheric water vapor feedbacks. The radiative forcing due to methane was computed using the linearized radiative efficiency from table 2.14 in Forster et al. (2007), amplified by feedback due to ozone and stratospheric water vapor effects for consistency with the calculation of GWP .

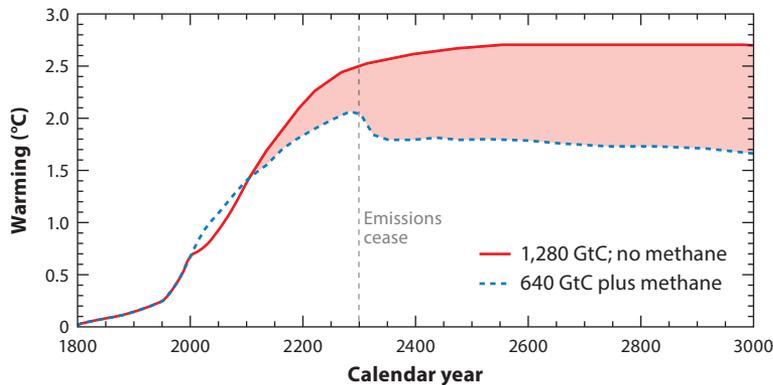


Figure 9

Warming resulting from the 1,280-GtC extended A2 CO₂ emission scenario of Eby et al. (2009), and from the 640-GtC extended A2 scenario augmented by methane emissions at a constant rate between the years 2000 and 2300 followed by zero methane emissions. The methane emission rate is chosen such that the GWP_{100} -weighted emissions in the two cases are identical.

The solid red curve corresponds to a scenario in which higher CO₂ emissions have been tolerated but methane has been abated. The dashed blue curve corresponds to a scenario in which the methane emissions have been left untouched, but instead a GWP_{100} -equivalent amount of CO₂ has been abated. So far as GWP_{100} is concerned, the two resulting climate futures should be equivalent, but it is clear that they are not. The methane abatement scenario does yield a very slightly cooler climate out to the year 2100, but this comes at the expense of a much warmer climate thereafter. Moreover, choosing methane abatement over CO₂ abatement results in an extra degree of warming, which persists to the end of the millennium and beyond. The calculation also illustrates the reversibility of the methane-induced warming, as is manifest in the sharp drop in temperature following the cessation of methane emissions in year 2300.

The chief failures of the GWP metric are summarized in the sidebar, Problems with Global Warming Potential. The final item of the list, concerning nonlinearity, can be fairly easily addressed. In the numerator of GWP , nonlinearity affects only CH₄ and N₂O, and even then only for unusually large concentration changes; if such effects become important, they can be addressed quasi-linearly by making the GWP weights dependent on prevailing concentrations at the beginning of each year. The more serious CO₂-related nonlinearities in the denominator of GWP can be addressed using the linearity between cumulative CO₂ emissions and radiative forcing (see Section 2), as proposed by Caldeira & Kasting (1993). The remaining failures, however, appear insurmountable if one's goal is to characterize the aggregate effect of short-lived and millennial greenhouse gases.

With such severe and evident shortcomings, how is it that GWP has become the de facto standard for analysis and implementation of multigas climate protection strategies? The challenge posed by aggregating CO₂ with shorter-lived gases was recognized from the start, and indeed GWP was initially introduced more as a means of illustrating the difficulties than as a proposed solution. In the IPCC First Assessment Report, GWP was introduced with the caveat, "A simple approach [i.e., the GWP] has been adopted here to illustrate the difficulties inherent in the concept," as noted by one of the originators of the concept (Shine 2009). Since that time, there has been no shortage of published work on the inadequacies of GWP , covering most of the issues raised in this section and a number of others besides (O'Neill 2000; Smith & Wigley 2000a,b; Shine et al.

PROBLEMS WITH GLOBAL WARMING POTENTIAL

- GWP_b is very sensitive to the time horizon b used in computing it.
- GWP_b is completely insensitive to climate changes that happen after time b , no matter how catastrophic they may be. A corollary of this property is that GWP_b doesn't distinguish between the reversibility of the effects of short-lived greenhouse gases and the essential irreversibility of the effects of millennial gases such as CO_2 .
- GWP_b is not suitable for use in conjunction with target temperature goals, whether expressed as temperature at a fixed date or peak temperature at whatever time the peak occurs. In particular, it exaggerates the contribution of early SLCP abatement to such goals, because it does not capture the “deferrability” of SLCP abatement.
- For emissions sustained over a time comparable to or longer than b , GWP_b greatly underestimates the warming due to CO_2 or other millennial gases, relative to that which would be caused by emission of short-lived gases with the same GWP_b -weighted value.
- GWP is based on pulse emissions, but the actual mean radiative forcing depends on the emissions trajectory of the gases under consideration. This leads to errors in the estimate of the climate impacts of gases whose emissions trajectories deviate substantially from a pulse, as is typically the case. The error is particularly serious when dealing with gases with very long lifetimes.
- As conventionally defined GWP neglects nonlinearity, both in the relation between radiative forcing in concentration and in the relation between emission and concentration.

2007, 2009; Manning & Reisinger 2011; Solomon et al. 2012). GWP seems to live on mainly as a result of inertia. It has little to recommend it beyond its simplicity, but its simplicity is of a sort that conforms to H.L. Mencken's dictum, “There is always a well-known solution to every human problem—neat, plausible, and wrong.”

3.4.2. Global temperature potential. Figures 7 and 8 show that GWP is a poor predictor of the temperature change that is the mediator of climate damages. In order to rectify this problem, Shine et al. (2005) proposed the global temperature change potential, which (like GWP) is defined relative to a time horizon b . We call this GTP_b . GTP_b is defined as the global mean warming at time $t_0 + b$ that would result from a pulse emission of a given mass of greenhouse gas at time t_0 , relative to the warming that would be caused by release of the same mass of CO_2 . Whereas GWP_b requires only that one be able to compute the radiative forcing due to a pulse emission, computation of GTP_b must involve a climate model, which can be anything from a simple model like Equation 4 up to a full general circulation model. When b is greater than the lifetime τ of the greenhouse gas in question, GTP_b will always be less than GWP_b . GTP does much better than GWP at characterizing the amount of climate change remaining a time b after emissions cease. It shares with GWP the problem of insensitivity to what happens after time b , and in particular doesn't capture the distinction between the reversibility of warming due to short-lived gases versus the essential irreversibility of CO_2 -induced warming.

If used with a fixed b , GTP_b weighting has some serious shortcomings when used in an attempt to achieve a target temperature goal. Suppose the goal is to limit warming in year t_1 to below some threshold value; t_1 might be a fixed year, or it might float in accordance with the timing of anticipated peak warming. For a gas with lifetime τ , the main contribution to warming at t_1 is from emissions occurring between $t_1 - \tau$ and t_1 . This contribution will be missed if b is held fixed, particularly if b is significantly larger than τ . To address this problem, Shine et al. (2007) introduced a dynamic variant of GTP . In essence, one reduces b as the target year approaches,

specifically weighting emissions in year t with the time-dependent weight GTP_{t_1-t} . This approach, in which short-lived emissions are accorded more value as the target year is approached, is similar to that used in the analysis of Manne & Richels (2001) but requires no explicit consideration of economic factors.

Shine et al. (2005) proposed another alternate metric, $GTPS_b$, which is based on the temperature response to sustained emissions (i.e., a fixed emissions *rate*) rather than a pulse release. $GTPS_b$ has a number of problematic aspects, noted in Shine et al. (2005), but the most serious shortcomings are that it fails to reflect the fact that early emissions contribute little to warming in the target year and that it shares with GWP_b the problem of underestimating the climate impact of CO₂ emissions that persist for longer than b . The use of pulse GTP_b with time-dependent b is better suited to implementation of target temperature goals.

3.4.3. Are metrics really necessary? In light of the results of Section 3.3, no metric that aggregates CO₂ emissions with emissions of short-lived gases can faithfully represent all aspects of the impact of emissions on climate. If such a metric is to be used, a decision must be made as to which aspects of climate are most important to get right and in which aspects serious distortions are to be tolerated. In this sense, the metric becomes a distillation of the set of values that inform climate protection goals. For example, as noted by O'Neill (2000), formulation of a metric involves decisions about how much one values prevention of short-term versus long-term or irreversible climate change. GWP and GTP are described in Bond et al. (2013) as “purely physical” metrics, but in fact they incorporate deeply embedded value judgments regarding the importance of short-term versus long-term effects. A metric can never be truly value free, so if some form of aggregation of climate effects is deemed to be necessary, the formulation of the metric should be seen as a job to be done transparently in light of society’s goals, for example, as part of the process by which a treaty is negotiated.

There is a particular danger that policymakers will view a metric as a statement of objective scientific fact, and not recognize the value judgments that have been buried in the metric’s definition. Though recent IPCC assessments (e.g., Forster et al. 2007) provide values for GWP_{20} , GWP_{100} , and GWP_{500} , policy assessments almost invariably adopt GWP_{100} , often without any statement of rationale and sometimes without even stating that a choice has been made, as in the CO₂-equivalent values used throughout IEA (2009). For the sake of transparency, it is best to use metrics as little as possible, if at all, in evaluating climate futures.

Metrics are never really needed for analysis of the climate consequences of emission control strategies and indeed are counterproductive because they obscure valuable information about climate forcing. Given a set of emissions trajectories of individual gases, the corresponding radiative forcing time series can be unambiguously computed, and in many cases this itself is sufficient for evaluating the effectiveness of various alternatives. This approach was taken in analyzing the consequences of future HFC emissions in Velders et al. (2009). When it is necessary to translate radiative forcing into global mean temperature change, simple models in the spirit of Equation 4 are generally sufficient. This is the approach taken by Cathles (2012) in analyzing the climate benefits obtained by substituting natural gas for coal in electric power production.

It is less easy to dispense with metrics in formulating climate protection protocols, especially emissions trading or tax schemes. One can begin by defining different metrics for different lifetime classes of emissions and adopt a multibasket approach analogous to that used in the Montreal Protocol (Daniel et al. 2011). However, if the protocol is implemented by assigning a monetary value to each individual metric-weighted emission, then the economic system will simply add up the monetary values of the individual emissions classes and in effect create a single-basket aggregate metric out of them. To some extent the problem can be addressed by phasing in the

Black carbon (BC):
solid particulate
aerosols composed
primarily of elemental
carbon; similar to soot

incentives for SLCP mitigation only after CO₂ emissions have fallen sufficiently far, but even then there is the risk that the phase-in of SLCP incentives could slow or reverse the decline in CO₂ emissions. Designing a protocol that achieves the desired behavior of both SLCP and CO₂ emissions is a great challenge, and one which we do not attempt to solve here. Lauder et al. (2013) have proposed a methane control approach that does not involve emission-equivalent metrics and explicitly recognizes the fact that for methane the climate change responds primarily to emissions rates rather than cumulative emissions. This approach can be extended to other forms of SLCP, and indeed Smith et al. (2012) have proposed a means of objectively formulating lifetime-specific metrics for use with multibasket policy vehicles, but the study of such strategies is still in its infancy.

4. BLACK CARBON

The term black carbon (BC) refers to solid particulate aerosols composed primarily of elemental carbon. BC affects the climate almost exclusively through absorption of solar radiation, which occurs both when the particles are airborne and when the particles become incorporated in surface snow and ice. BC (also known as soot) particulates have long been studied in the air pollution community because of their respiratory health effects, but in the past decade interest in BC as a climate forcing agent has skyrocketed. Here we highlight only a few key BC results germane to the main issues treated in the present review; the reader seeking an in-depth understanding of BC climate effects is referred to the thorough assessment that has recently appeared (Bond et al. 2013).

Like all aerosols, the atmospheric lifetime of BC is extremely short, ranging from 3.3 to 10.6 days (Bond et al. 2013, section 5.2), with removal primarily due to the scavenging by precipitation that becomes possible once fresh BC particles become coated with liquids that allow them to be more easily incorporated in cloud droplets. Because BC is readily combustible at high temperatures, efficient controlled combustion such as occurs in large-scale coal-fired power plants is not a significant source of BC. Most BC comes from open burning of biomass and inefficient combustion of coal such as occurs in domestic heating and small-scale industrial processes. Burning of diesel fuel in engines is also a significant source of BC.

The very short lifetime of BC means that whatever adverse climate impacts it may have, the issues regarding BC versus CO₂ mitigation are essentially identical to those we have already treated in the context of well-mixed short-lived greenhouse gases such as methane and HFCs. Notably, the issue of reversibility of BC effects versus irreversibility of CO₂ effects, and the closely related implications of scaling of climate forcing with emissions rate for short-lived substances versus cumulative emissions for CO₂ and other millennial gases, apply to BC in the same fashion as in our discussion above. The chief novel difficulty presented by BC is quantifying the resulting climate damages in a way that allows the effect of BC emissions to be compared with the effect of CO₂ emissions. Many assessments simply restrict attention to BC effects that can be characterized by the influence of BC on global mean warming, and that is the approach we generally adopt here, though we also highlight reasons this may give an incomplete picture.

4.1. Radiative Effects of Airborne Black Carbon

As its name suggests, black carbon is a potent absorber of solar radiation. Sunlight passing through a mere 0.09 g of fresh BC particles suspended in an atmospheric column with a 1-m² base will be attenuated by a factor of 2, leaving behind half of its energy in the column (Bond et al. 2013, section 2.1). Coating of particles by a small amount of transparent aerosol-forming liquid as the particles age increases the strength of absorption by about 50%, though if the coating becomes massive enough to dominate the BC content, the resulting aerosol can become predominantly

reflecting. For anthropogenic BC, typical paths in the atmospheric column are on the order of 10^{-3} g/m² or less averaged along a latitude circle. Most of the mass is concentrated near the ground and in the Northern Hemisphere tropics. This would lead to absorption of only 0.75% of the incident sunlight, and even this figure overstates the influence on the top-of-atmosphere radiation budget because, over a dark surface like the ocean, most of that sunlight would have been absorbed at the ground anyway. Still, the small proportion absorbed, when applied to 300 W/m² of daily averaged incident sunlight, amounts to 2.25 W/m² added to the atmospheric heating, at the expense of a somewhat lesser amount taken away from absorption at the ground.

Because BC particles are much smaller than the wavelength of infrared light, they absorb and emit infrared inefficiently. Hence, with the typical anthropogenic BC burden, BC has no significant effect on Earth's infrared radiation budget, particularly in view of the fact that BC is concentrated in the lower atmosphere, which radiates at a temperature similar to that of the ground. Infrared effects could become significant if the atmospheric burden became substantially greater, particularly at high altitudes, but this is unlikely to occur outside of extreme climates such as those engendered by nuclear winter (Cess et al. 1985) or post-asteroid impact scenarios.

4.2. Effects of Black Carbon on Climate

The climate response to BC aerosols is very different from that to well-mixed greenhouse gases because the very short lifetime of the aerosols leads to radiative forcing that is geographically and seasonally highly inhomogeneous. This is also the case for reflecting aerosols such as sulfates, but BC presents the even more severe challenge that its chief effect is to change the vertical distribution of deposition of solar energy in the Earth system, depositing directly in the atmosphere solar energy that would otherwise be absorbed at the ground and communicated to the atmosphere by convection. Because the combined effect of scattering of sunlight from the atmosphere, clouds, and the underlying surface is reduced by addition of BC, BC does result in a net positive top-of-troposphere radiative forcing, by an amount that increases in the presence of low clouds or a high-albedo surface such as ice or snow. This does generally (though not invariably) lead to surface warming, but the concomitant reduction in surface solar absorption and change in altitude of deposition greatly complicates the range of possible climate responses to BC forcing.

Because surface insolation tends to warm Earth's surface relative to the temperature of the overlying atmosphere (Pierrehumbert 2010, chapter 6), the reduction in surface insolation caused by BC can under some circumstances cause the surface to cool at the same time that the atmosphere warms. This phenomenon is familiar from the extreme case of nuclear winter, in which the prime effect of atmospheric soot is a catastrophic cooling of the surface (Cess et al. 1985), but it can also result from the more modest BC burdens resulting from less extreme anthropogenic emissions. In the tropics, this effect can be seen over the Indian subcontinent in the simulations of Meehl et al. (2008). In the arctic, Flanner (2013) has shown that high-altitude BC can cool the surface, and Ban-Weiss et al. (2012) have found that surface cooling is a general consequence of high-altitude BC, even at fairly low BC burdens.

The reduction of surface insolation leads to a more robust effect on precipitation, because precipitation is largely controlled by the surface energy budget; the energy needed to sustain evaporation is supplied mainly by surface insolation, especially in the tropics, so that a reduction in surface insolation tends to reduce precipitation. This effect can be seen clearly in the analysis of Ming et al. (2010), though Ban-Weiss et al. (2012) have found that there are nonetheless circumstances in which low-level BC can increase precipitation.

Climate change is more complex than just global mean warming, but for well-mixed greenhouse gases the pattern of climate effects scales reasonably well with the global mean warming; a degree of

warming due to methane entails pretty much the same constellation of regional effects as a degree of warming due to CO₂. This simple scaling does not apply to very inhomogeneous forcings, and is yet more problematic for BC, which alters the locus of deposition of solar energy. The implications of the short-term nature of the climate changes due to BC still mean that the issues regarding priority of BC versus CO₂ mitigation are much the same as for methane or HFCs. However, because the climate damages of BC cannot be indexed to global mean warming, translating BC emissions into damages in a way that can be compared with CO₂ or other well-mixed greenhouse gases would require a sophisticated damage model, subject to many uncertainties as to how the damages should be quantified. However that is done, it remains true for BC, as for methane or HFCs, that it is the emission *rate* rather than cumulative emissions that drives damages.

4.3. The Importance of Reflecting Coemittants

The abstract of Bond et al. (2013) contains the easily quotable summary statement: “We estimate that black carbon, with a total climate forcing of +1.1 W m⁻², is the second most important human emission in terms of its climate forcing in the present-day atmosphere; only carbon dioxide is estimated to have a greater forcing.” This statement, taken out of context, easily leads to the misconception that BC mitigation offers climate benefits superior even to abatement of methane, which is the number two well-mixed greenhouse gas behind CO₂. This interpretation, however, neglects the important qualifications stated equally prominently elsewhere in the abstract. The radiative forcing quoted above is a fictitious quantity corresponding to the radiative forcing that would occur if BC sources emitted BC alone, and not other aerosol-producing substances. In reality, all BC sources also emit precursors to reflecting aerosols that have a cooling effect on climate both through direct scattering and through indirect effects on clouds. Sulfur-rich fuels produce sulfate aerosols, and burning of all forms of biomass produces precursors to secondary organic aerosols. Uncertainties in the estimate of the indirect aerosol effect, however, lead to large uncertainties in the net climate forcing due to all industrial-era BC-rich sources. The best estimate for the net climate effect is a slight cooling (−0.06 W/m²), with an uncertainty range spanning −1.45 to +1.29 W/m² (Bond et al. 2013). Thus, it is far from clear that across-the-board mitigation of BC sources would cool the climate, and if anything it is more probable that some warming would result. Therefore, moderation of global warming cannot be seen as a reliable benefit of BC mitigation. At best, aggressive mitigation of BC sources such as diesel engines, which produce relatively little reflecting aerosol, can be seen as a way to partially offset the warming caused by mitigation of biomass burning, which is likely to be necessary on the grounds of human health effects. It is possible that regional effects, and other climate impacts that are not accompanied by global mean warming (Section 4.2), could nonetheless count as climate benefits of BC mitigation, but it is hard to quantify these benefits and still harder to express them in terms of an equivalent in mitigation of well-mixed greenhouse gases, even if one skirts the lifetime issue by restricting the comparison to short-lived gases such as methane.

5. ANALYSIS OF MITIGATION STRATEGIES

In this section we use the general principles developed in the preceding sections to address a number of concrete questions concerning the consequences of various mitigation strategies involving short-lived climate pollution. Climate stabilization requires that CO₂ emissions go to essentially zero at some point; otherwise Earth will keep warming indefinitely (Matthews & Caldeira 2008), in which case SLCP mitigation will be of little consequence. Therefore, all the scenarios discussed here presume that CO₂ emissions cease completely at some point, though with varying

amounts of cumulative carbon having been emitted by the time of cessation. Specifically, for CO₂ emissions and corresponding atmospheric concentrations, we make use of the RCP scenarios as extended by Zickfeld et al. (2013) to the zero-emissions point and beyond that to the year 3000, with emissions diagnosed from the UVic model as described in that paper. The RCP scenarios are labeled according to the maximum radiative forcing encountered by year 2100, so larger numbers correspond to less constrained emissions.

We characterize SLCP emissions by the radiative forcing they cause. The SLCP-associated radiative forcing in year 2000 is taken from figure 2.20 of Forster et al. (2007), with some allowance for efficacy factors. An idealized future SLCP scenario is constructed as follows. SLCP radiative forcing is assumed to grow linearly to a maximum value in year 2100 and thereafter remain constant. Methane radiative forcing is assumed to grow by 0.25 W/m². This is midway between the RCP6 and RCP8.5 scenarios; note that the very high methane emissions growth in RCP8.5 is realized only as a result of extremely high fossil fuel usage. This radiative forcing is amplified by a factor of 1.45 to account for ozone and stratospheric water vapor feedbacks and the radiative efficacy factor. HFC radiative forcing grows to 0.2 W/m², which is somewhat higher than the RCP8.5 value but lower than implied by the highest estimate in Velders et al. (2009). N₂O goes up by 0.29 W/m² (midway between RCP6 and RCP8.5), while Montreal gases go down by 0.31 W/m². We do not attempt to estimate a trend in either BC or ozone radiative forcing in the baseline scenario, apart from ozone changes implied as part of the methane feedback. This adds up to a growth by 0.54 W/m², starting from an initial value of about 1.5 W/m².

It is impossible to avoid simultaneous consideration of the negative radiative forcing due to sulfate aerosols, because without taking them into account the present climate is considerably warmer than observed if IPCC midrange climate sensitivity is applied. A good fit to observations can be obtained if it is assumed that present aerosol cooling cancels out the SLCP radiative forcing (Eby et al. 2009); we take this as the starting point in the aerosol scenario. Future sulfate emissions should be tied to fossil fuel usage in the various RCP scenarios, but because we wish to emphasize the implications of various SLCP mitigation strategies here without being distracted by differences in reflective aerosol forcing, we adopt a simple scenario in which the aerosol cooling is held at a constant negative value until year 2100 and thereafter increases linearly to zero by the year 2200 as fossil fuel emissions wind down to nearly zero. The ultimate and inevitable loss of aerosol cooling that accompanies termination of fossil fuel usage leads to additional committed warming due to the unmasking of greenhouse gas forcing (Ramanathan & Feng 2008, Matthews & Weaver 2010). Both the current sulfate aerosol forcing and the climate sensitivity are uncertain, and for consistency with the instrumental climate record the uncertainties need to be considered jointly, as done in Armour & Roe (2011); we do not pursue uncertainty analysis here. The reference scenarios for SLCP and sulfate aerosol radiative forcing are shown in **Figure 10**.

We consider two SLCP mitigation scenarios, shown in **Figure 10**. The amount of abatement is measured in terms of the net reduction in radiative forcing achieved. In each scenario, the mitigation starts at zero in the initial year, becomes linearly more negative for the first 100 years, and remains constant thereafter. In the early abatement scenario, the SLCP mitigation begins immediately, whereas in the late abatement scenario the start is deferred by 100 years. We estimate the maximum SLCP radiative forcing available for abatement as follows. Present-day ozone offers 0.3 W/m², HFCs offer 0.2 W/m² by 2100, and methane could grow to 0.75 W/m² (which we augment by a factor 1.45). It is hard to know what to allow for possible BC mitigation, because the best estimate of present emissions taking into account committed reflecting species yields a slightly negative radiative forcing, with a range from -1.45 to +1.29 W/m²; we assume 0.5 W/m² is available for mitigation, which is partway toward the upper extreme of the uncertainty range. This adds up to about 2 W/m², and we assume that half of that can actually be mitigated. The

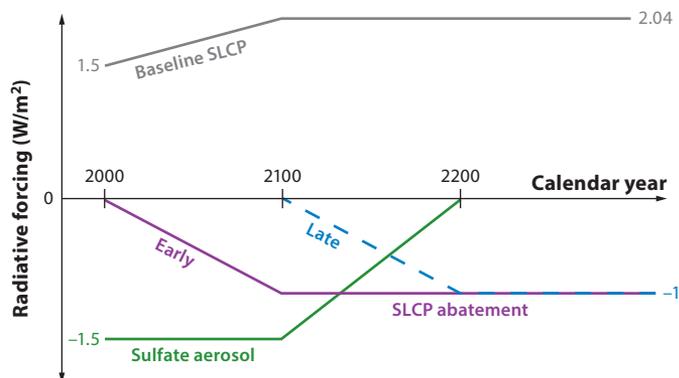


Figure 10

Baseline short-lived climate pollution (SLCP) scenario, with two SLCP mitigation options. The numbers with the curves indicate values of radiative forcing in W/m^2 .

assumed amount of early SLCP mitigation is somewhat greater than that in UNEP (2011) but is spread out over 100 years rather than concentrated in the first 50. Smith & Mizrahi (2014) argue that the potential for abatement and the resulting climate benefits are both greatly overestimated in UNEP (2011), but we adopt a strong abatement scenario here to highlight that the benefits of early abatement are slight even if one makes an optimistic assumption about the amount of feasible abatement. Our SLCP scenario is also biased favorable to the case for SLCP abatement in that we do not subtract off anything for the considerable amount of methane abatement that would be obtained as an automatic cobenefit of reducing fossil fuel production.

The two-box model temperature simulations in this section were carried out with a reduced deep-ocean heat uptake coefficient $\gamma = \frac{1}{2}\lambda$ instead of the default value, because the reduced value more closely mimics the results of other simplified climate models used in policy analysis.

5.1. Does Early SLCP Mitigation “Buy Time” to Implement CO_2 Mitigation?

Advocates of early SLCP mitigation almost invariably acknowledge that CO_2 mitigation is an indispensable part of any climate protection strategy. However, recognizing that the prospects for serious CO_2 mitigation appear dim in the near future, the idea has sometimes been advanced that early SLCP mitigation (which is perceived, rightly or wrongly, as more feasible than early CO_2 mitigation) could in some way “buy time” and help hold back global warming while we are waiting for a more enlightened era to implement the required CO_2 mitigation.

The notion of “buying time” can be made precise by restating it in the following way. Suppose that there is a CO_2 emissions trajectory $E_0(t)$ that allows the agreed-upon climate targets to be met, when supplemented by *eventual* SLCP mitigation applied in a century or so, once CO_2 emissions are trending toward zero. Now suppose instead that, in recognition that the actions needed to keep near-term CO_2 emissions from exceeding E_0 are not likely to be put in place, early SLCP mitigation is implemented in an attempt to make up for the lack of CO_2 mitigation. Given early instead of late SLCP mitigation, what modified CO_2 emissions trajectory $E_0(t) + \Delta E(t)$ achieves precisely the same radiative forcing $\Delta F(t)$ as formerly would have been achieved with late SLCP mitigation if the target E_0 had been met? The CO_2 emissions modification $\Delta E(t)$ characterizes the additional time “bought” by implementing early SLCP mitigation.

Let $\Delta F'(t)$ be the radiative forcing due to the difference between early and late SLCP mitigation. For the idealized mitigation ramp described above, it takes the form shown in **Figure 11a**. The

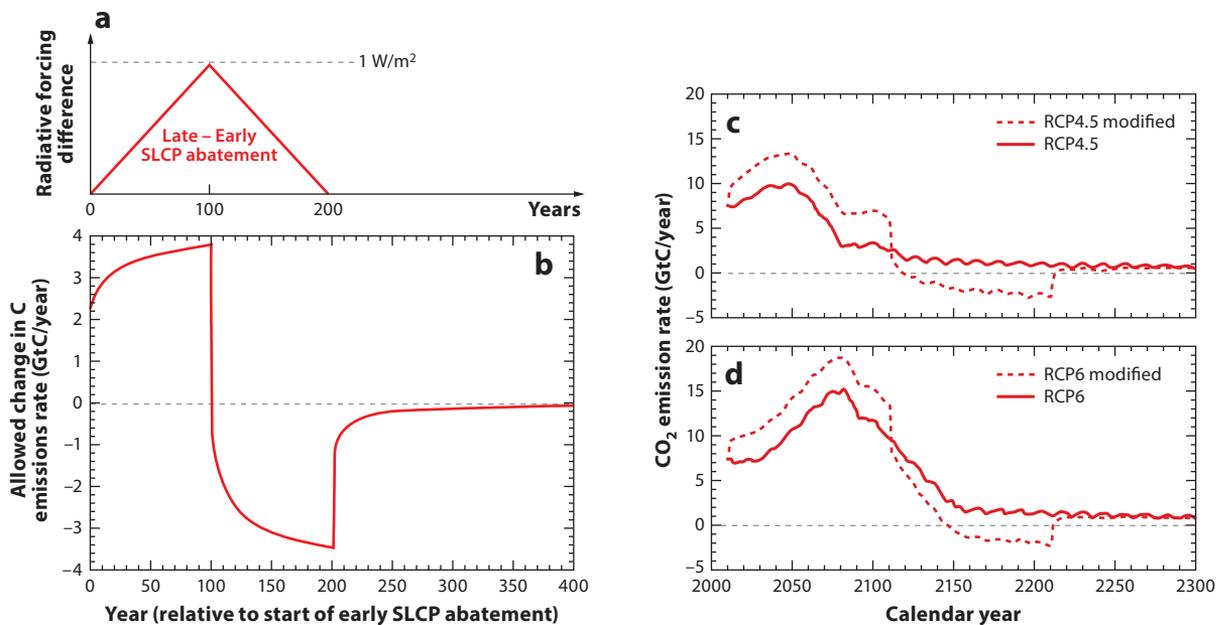


Figure 11

(a) The difference in radiative forcing between early and late short-lived climate pollution (SLCP) mitigation scenarios. (b) The change in CO₂ emissions rate that is allowed as a result of early SLCP mitigation, subject to the requirement that net radiative forcing versus time is unchanged. (c,d) Modified RCP CO₂ emissions scenarios obtained by adding the emission perturbation in panel b to the (c) RCP4.5 and (d) RCP6 scenarios.

corresponding CO₂ emission change can be obtained by then solving Equation 3 for $\Delta E(t)$. The results are shown in **Figure 11b**. It is seen that indeed the early SLCP mitigation allows extra CO₂ emissions in the first century. However, these extra emissions must be paid for with additional steep emissions reductions in the century following. This is a simple consequence of the fact that radiative forcing for CO₂ is primarily determined by cumulative carbon emissions. The extra CO₂ abatement in the second century must be done *on top of* whatever abatement measures were already required to achieve E_0 . Thus, early SLCP mitigation doesn't "buy time" to implement the measures that were initially thought necessary—one needs to make up for lost time in the first century by implementing all the deferred CO₂ mitigation measures that were not done then.

To provide a clear idea of the kind of emissions trajectory one would need to implement in order to take advantage of the time "bought" by early SLCP mitigation, in **Figure 11c** we show the results of adding ΔE to the CO₂ emissions trajectory in the RCP4.5 scenario. This scenario represents an only marginally satisfactory climate target, as it yields a peak warming of 3°C (Zickfeld et al. 2013). Even with such a mild target, fully taking advantage of the flexibility represented by ΔE requires emissions to fall to zero by the year 2120, and actually requires *negative* emissions (i.e., large-scale implementation of air capture) for the century thereafter. Compared to a situation in which CO₂ mitigation is phased in earlier and more gradually, this does not look like a particularly feasible climate stabilization strategy. Even the higher-emission RCP6 scenario, which yields an alarming 4.2°C of warming, would require a precipitous decline in CO₂ after 2100, followed by negative emissions (**Figure 11d**).

Using a different approach, Allen & Stocker (2014) have also analyzed the adverse consequences of any delay in CO₂ mitigation. Their analysis also demonstrates that there is an inherent risk in

early SLCP abatement, in that the climate protection value of such abatement is not known until it is known when CO₂ emissions actually peak and how rapidly the emissions can be brought to zero; overoptimistic assumptions about CO₂ mitigation lead to overestimates of the importance of early SLCP abatement. This argument supports strategies in which SLCP abatement is delayed at least until CO₂ emissions have begun to decline.

5.2. Does Early SLCP Mitigation Help Meet Peak Warming Targets?

In this section we show that even if early SLCP mitigation is implemented without modifying CO₂ emissions, it has no effect on the peak warming and therefore does not help to meet peak warming targets. This result holds simply because the peak CO₂-induced warming occurs near or somewhat after the time CO₂ emissions approach zero (Section 2), whereas the warming due to SLCP emissions more than a decade or two before the peak decays so quickly that it doesn't contribute materially to the peak value (Section 3.3).

Figure 12 shows the warming resulting from the RCP4.5, RCP6, and RCP8.5 CO₂ emissions, augmented by baseline SLCP and sulfate aerosol trajectories plus either early or late SLCP mitigation. It is seen that early SLCP mitigation has some value if it can be done without causing an increase in CO₂ emissions, in the sense that it yields a somewhat cooler climate in the first century than would otherwise be obtained. Implementing early SLCP mitigation delays the time of crossing a 2°C threshold by 68 years for RCP4.5, but only 17 years for RCP6 and 9 years for RCP8.5; if measures are not taken to drastically reduce CO₂ emissions, the rapidly accumulating warming due to CO₂ quickly eats up any benefits of early SLCP mitigation.

This calculation shows that SLCP mitigation is most significant only if measures are in place to put the world on a low-CO₂ emissions trajectory; otherwise, the contribution to be made by SLCP appears marginal in comparison to the CO₂ effect. In a world as hot as that yielded by RCP8.5, things would be so bad that the slight additional near-term warming caused by deferring SLCP mitigation would not rate high on our list of concerns. In other words, the first order of business is to get the world onto an RCP4.5 or lower CO₂ emissions trajectory, and in comparison with that the consequences of early versus late SLCP mitigation appear minor.

In any event, if the prime climate protection goal is to limit peak warming, then early SLCP mitigation is pointless, because in no case does early SLCP mitigation significantly reduce the peak warming. The calculation does show, however, that *eventual* SLCP mitigation helps trim the magnitude of the peak warming.

Bowerman et al. (2013), using a somewhat different set of assumptions, also arrived at the conclusion that doing SLCP mitigation early rather than late has no effect on peak warming. They conclude further that under most circumstances of interest, accelerating SLCP mitigation does not reduce even the maximum *rate* of warming.

5.3. What Happens if SLCP Mitigation Substitutes for Some CO₂ Mitigation?

The above examples assume that the SLCP mitigation was gotten “for free,” in the strong sense that introduction of SLCP mitigation (whether early or late) has no effect whatsoever on the amount of CO₂ mitigated. This is almost certainly an unrealistic assumption, as there are any number of ways that efforts to mitigate SLCP could displace CO₂ mitigation. Mitigation generally costs something, and any money spent on SLCP mitigation is money that is not available to be spent on CO₂ mitigation. Even if SLCP can be gotten at zero or even negative cost, as might be the case for fugitive emissions from natural gas production, any emissions trading or tax scheme that is based on a weighted equivalence of SLCP and CO₂ emissions would provide incentives for the maximum possible SLCP mitigation and minimum possible compensating CO₂ mitigation.

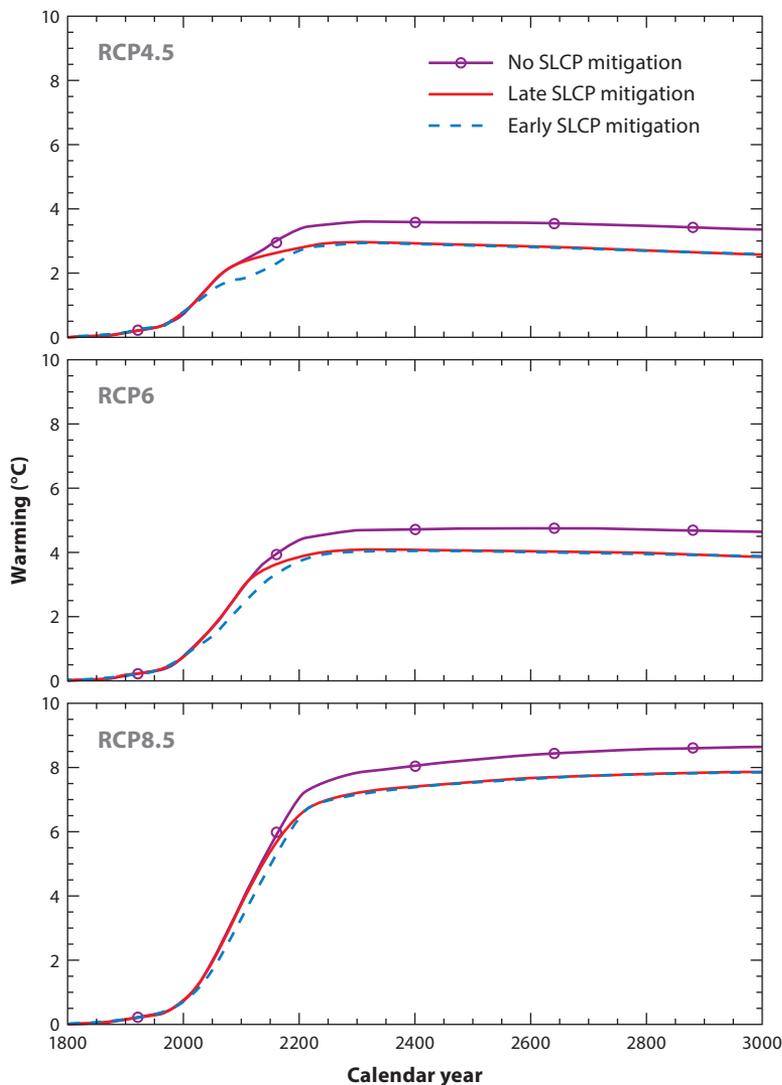


Figure 12

Comparison of warming from early, late, and no short-lived climate pollution (SLCP) abatement scenarios applied on top of a baseline consisting of RCP4.5, RCP6, and RCP8.5 CO₂ emissions plus the baseline SLCP and sulfate aerosol shown in **Figure 10**.

Moreover, side effects of SLCP mitigation could result in additional CO₂ emissions. For example, if premature introduction of HFC substitutes reduces the efficiency of refrigeration and air conditioning, or makes foam building insulation more expensive, the reduced energy efficiency will lead to more CO₂ emissions. Similarly, controls on BC from diesel could lead to less use of efficient diesel engines, or regulation of methane emissions from hydrofractured natural gas production could slow the replacement of coal with natural gas, either of which would tend to increase CO₂ emissions. Finally, there is the more diffuse issue of political capital, in the sense that any effort expended in building consensus for SLCP mitigation is effort that is not spent building consensus for CO₂ mitigation; to the extent SLCP efforts are successful, they may fulfill a sense

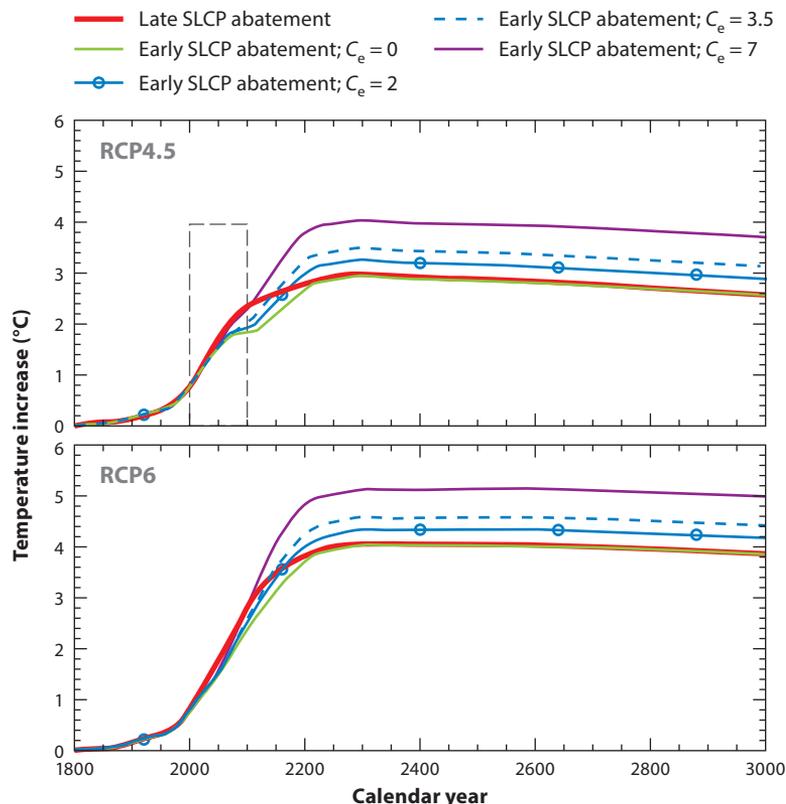


Figure 13

Effect of substitution of early short-lived climate pollution (SLCP) mitigation for CO₂ mitigation, at a rate determined by the equivalence factor C_e . $C_e = 0$ corresponds to the no-substitution early SLCP mitigation curve in **Figure 12**, and $C_e = 3.5$ is approximately equivalent to weighting by GWP_{100} . The dashed box includes the limited time range considered in UNEP (2011).

of having “done something” about climate, detracting from the sense of obligation to tackle CO₂ emissions.

In **Figure 9** we examined the effect of substitution of SLCP for CO₂ mitigation based on GWP_{100} equivalence. In **Figure 13** we consider the problem more generally. We formulate the trade-off between SLCP and CO₂ in terms of an equivalent emissions weight C_e analogous to GWP , but instead of expressing the equivalence as GtC of CO₂ emissions per Gt of SLCP emissions, we express it in terms of the radiative forcing due to the SLCP, i.e., as annual GtC emissions considered equivalent to 1 W/m² of SLCP radiative forcing. This still yields an emission-equivalent metric, because the radiative forcing due to a short-lived substance is proportional to the emission rate (see Section 3.3). Formulated in this way, substitution can be studied without introducing different equivalence factors for the different emissions that make up the SLCP. An equivalence factor of 3.5 GtC/year per W/m² of SLCP abatement is the same as using GWP_{100} weighting, using Equation 2 to determine CO₂ radiative forcing (or slightly less using formulae in Forster et al. 2007). We assume for the moment that the late SLCP abatement indicated in **Figure 10** can somehow be obtained without being paid for in excess CO₂ emissions, so that the excess CO₂ emissions are computed by applying the equivalence factor only to the radiative forcing difference between early and late SLCP abatement, shown in **Figure 11**.

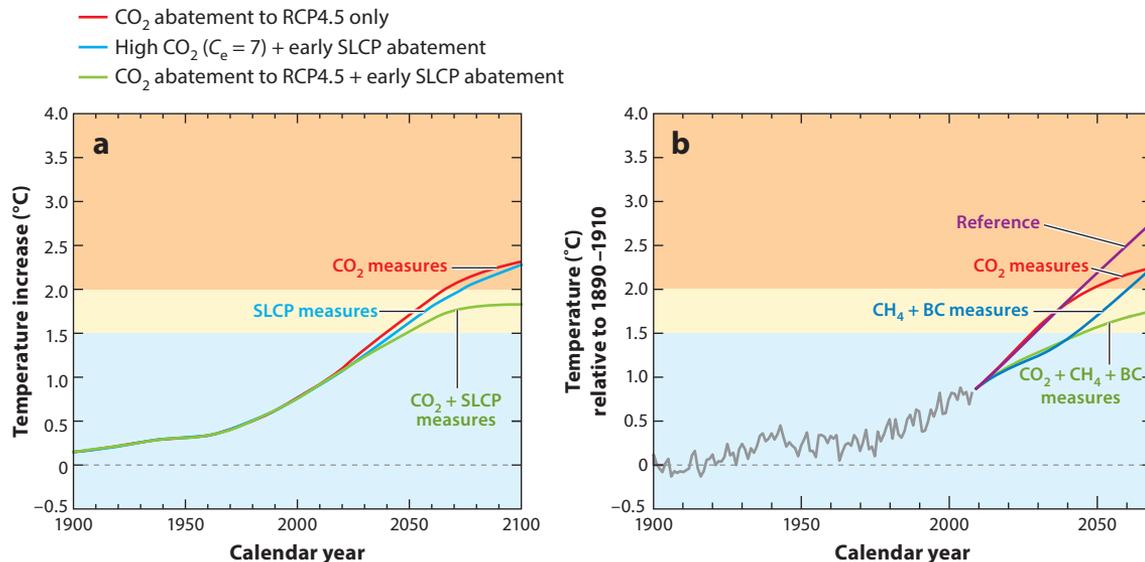


Figure 14

Comparison of (a) the portion the RCP4.5 case in **Figure 13** within the small dashed box with (b) figure ES-2 of UNEP (2011). Abbreviations: BC, black carbon; SLCP, short-lived climate pollution.

Results are shown in **Figure 13**. It is seen that *any* amount of substitution of SLCP abatement for CO₂ abatement results in a situation in which a temporary and modest reduction in near-term temperature is bought at the expense of a permanent increase in the long-term temperature. The greater the substitution, the less the short-term climate benefits and the greater the irreversible harm. Thus, the *only* early SLCP mitigation strategy that works is one that causes substantial SLCP mitigation without displacing *any* CO₂ mitigation that would otherwise occur, and even then only if the world is already on a low CO₂ emission pathway. For RCP6, the short-term benefits appear far less consequential, whereas the long-term harm remains evident.

One can easily get an exaggerated impression of the benefits of early SLCP mitigation if one looks at an overly narrow time range. Figure ES-2 of UNEP (2011), reproduced in **Figure 14b**, is a case in point. Looking out only to the year 2070, one would conclude that it is a matter of convenience whether to choose CO₂ mitigation or SLCP mitigation, with maybe some advantage in the latter. **Figure 14a** shows that, similarly, if one were to look only at the early portion of **Figure 13**, one would conclude that SLCP mitigation with C_e = 7 (which corresponds roughly to the CO₂ versus SLCP tradeoff examined in UNEP 2011) would accomplish almost exactly the same climate goals as CO₂ mitigation that achieved the RCP4.5 CO₂ emission trajectory. However, as one can see from the full results in **Figure 13**, the early SLCP strategy leads to a far warmer climate in the long run. Similar points about the dangers of assuming the world ends in 2070 have been made in Solomon et al. (2012) and Shoemaker & Schrag (2013).

5.4. Can the Value of Eventual SLCP Mitigation Be Translated into an Equivalent in Mitigation of CO₂ Emissions?

The above analysis of substitution of SLCP for CO₂ mitigation is rather generous toward SLCP mitigation in that it assumes that only the switch to early SLCP mitigation competes with CO₂

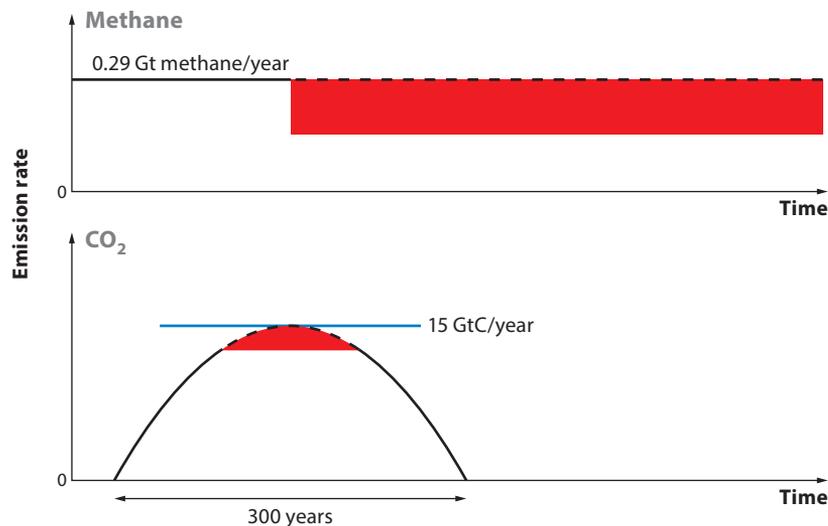


Figure 15

CO₂ mitigation (*lower panel*) that yields the same reduction in radiative forcing as a permanent 50% reduction in methane emissions rates relative to the current value (*upper panel*). Shaded red regions correspond to the mass of the gas mitigated, which in the case of methane is assumed to continue past the time range shown in the graph.

mitigation, and that late SLCP mitigation does not cause any similar displacement. If the 1 W/m² of eventual SLCP mitigation substituted for CO₂ mitigation in a manner similar to what we have assumed for early mitigation, then it would result in allowing an additional C_e GtC/year of CO₂ emissions to continue *forever*, leading temperatures to continue increasing indefinitely, until fossil fuels eventually run out (Matthews & Caldeira 2008). Yet, we have seen that once CO₂ emissions have dropped to nearly zero, mitigation of SLCP can make a valuable contribution to reducing peak warming (**Figure 12**). It is clearly beneficial to implement incentives for such mitigation to occur. But any protocol which treats X Gt of SLCP emissions as the equivalent of 1 GtC of CO₂ emissions will have adverse consequences.

The root of the problem is that any form of emission-equivalent weighting fails to capture the essential fact that radiative forcing for a short-lived substance like methane is proportional to the emission *rate*, whereas that from CO₂ is proportional to *cumulative emissions* (Section 3.3). This point has also been highlighted by Smith et al. (2012). It takes a *permanent* reduction in methane emission rate—corresponding to an infinite mass of methane abated—to yield the same long-term radiative forcing as abatement of a *finite mass* of CO₂, as illustrated in **Figure 15**. Specifically, using Equation 2 we find that a permanent reduction of SLCP emission rate corresponding to 1 W/m² is equivalent to a reduction of cumulative carbon emissions by 407 GtC, with regard to long-term radiative forcing. The example shown in **Figure 15** is scaled to a permanent reduction of methane emissions rate to half its present value.

A novel approach to multigas climate protection protocols, quite different from that used in the Kyoto Protocol, is required to properly deal with SLCP. In the context of a carbon tax, an emitter would pay a tax for each GtC of CO₂ emitted but would be given a one-time tax credit for each Gt/year of methane emissions *rate* reduction, weighted according to the corresponding radiative forcing. If the emitter ever increased the methane emissions rate again, the tax credit would need to be paid back with interest; moreover, for best results, the tax credit should be phased in only

after worldwide CO₂ emissions are trending downward at a sufficient rate. Related approaches to SLCP mitigation are discussed in Lauder et al. (2013).

6. CONCLUSIONS

On timescales relevant to climate policy decisions, the essential difference between short-lived forcing agents like methane, HFCs, or BC and millennial gases like CO₂ is that for short-lived agents temperature depends on the instantaneous emissions *rate*, whereas for millennial agents it depends on *cumulative emissions*. No climate policy that fails to respect this essential difference can have a good outcome. The key distinction can also be phrased in terms of reversibility versus irreversibility. The reversibility of the effects of short-lived climate pollution does not mean that early mitigation yields a large payoff in climate benefits. On the contrary, the reversibility means that their mitigation can be deferred without much harm, whereas deferring mitigation of a millennial gas commits Earth to an additional warming that is essentially irreversible.

For well-mixed decadal greenhouse gases (principally methane and HFCs), SLCP mitigation makes a significant contribution to climate protection only in a scenario in which cumulative CO₂ emissions are kept to a sufficiently low value. In such circumstances, *eventual* mitigation of SLCP emissions can yield a useful reduction in peak warming. However, *early* SLCP mitigation confers no advantage in meeting peak warming targets and will be harmful if it substitutes to any significant extent for measures that could have been deployed to reduce CO₂ emissions. Though early SLCP mitigation does not help reduce peak warming, it can moderate short-term warming somewhat, but this comes at the cost of considerably worse long-term warming if the implementation of SLCP measures substitutes to any significant degree for CO₂ mitigation. The only early SLCP mitigation strategy that works is one that causes substantial SLCP mitigation without displacing *any* CO₂ mitigation that would otherwise occur, and even then only if the world is already on a low-CO₂ emission pathway.

A common fallacy in assessing the value of SLCP mitigation is to think that because SLCP radiative forcing is nearly half the current total anthropogenic radiative forcing, eliminating SLCP emissions would solve nearly half the global warming problem. This reasoning would be valid if CO₂ emissions were to drop to zero in the next few years, but because CO₂ accumulates in the atmosphere, unlike SLCP, the actual future emissions add radiative forcing on top of the existing radiative forcing, and do so at an exponentially growing rate until CO₂ emissions are brought under control. SLCP radiative forcing can grow only to the extent that SLCP emissions rates themselves grow. Thus, the further out one looks, the more CO₂ dominates the climate problem, particularly in situations in which CO₂ mitigation remains inadequate.

To the extent that BC emissions cause warming, the issues are largely the same as for decadal greenhouse gases. The climate effects of BC are at present highly uncertain; when one takes into account the cooling effects of coemitted species, the best estimate of the net effect of BC-rich sources is a slight net cooling. BC has a range of effects on regional climate, notably precipitation, that are not captured by its radiative forcing alone, but any harm due to these effects is tied to instantaneous emission rates and therefore engages the same issues as were discussed for the decadal greenhouse gases. However, BC-rich sources have a profound adverse effect on human health. The best rationale for early abatement of BC emissions is that stated in Bond et al. (2013): “Reduction in aerosol forcing by mitigating BC-rich sources would be accompanied by small to no changes in short-term climate forcing . . . so mitigation of most BC-rich source categories offers a method of reducing health impacts with lower risk of adverse effects on climate” (Bond et al. 2013, section 10). Emphasis on the highly uncertain climate benefit is only a distraction, and it increases the risk that BC abatement could displace more pressing CO₂ abatement. If BC is a

major threat to human health, it makes no sense to target abatement only to those sources that cause the greatest net climate benefit, any more than it would make sense for China or India to defer much-needed efforts at reducing sulfate pollution on the grounds that this will lead to more warming. BC abatement is more of a piece with all those other things that need to be done to improve human welfare—e.g., reduce mercury emissions or improve access to clean drinking water—even though such things cost money, effort, and political capital and may in that sense compete for resources with climate protection. The climate impacts of tropospheric ozone are more certain than those of BC, but ozone is similar to BC in that mitigation can be amply justified on the basis of health consequences alone. But overemphasizing the climate cobenefits of ozone abatement is a distraction and increases the risk that such efforts will displace CO₂ mitigation.

As a means of aggregating the climate effects of short-lived climate forcers with those of millennial gases like CO₂, the widely used global warming potential (*GWP*) metric is hopelessly flawed, and the time has come to retire it. It is not needed for analysis of climate policies, and it does not provide a suitable basis for regulation, emissions trading, or emissions taxes. However, something very like *GWP* is useful for aggregating all the climate effects of substances with decadal or shorter lifetimes. Specifically, the quantity $a'\tau$, where a' is the radiative efficiency of the substance per unit mass of emission and τ is its lifetime, characterizes the climate impact of the emission (Section 3.3). If normalized to the value of $a'\tau$ for methane, it yields a methane-equivalent weighting. In using this metric, one still needs to bear in mind that for decadal and shorter lifetimes, it is the methane-equivalent *rate* that determines the climate impact.

For CO₂, cumulative carbon provides a completely satisfactory metric of climate impact. Other millennial gases, such as SF₆ and PFCs, can easily be incorporated into the metric as equivalent cumulative carbon, because their accumulated radiative forcing is proportional to their cumulative emissions and can be translated into the amount of CO₂ emissions that would cause the same radiative forcing using Equation 2.

We are not arguing that SLCP mitigation is irrelevant, only that it should take a backseat to CO₂ mitigation at least until CO₂ emission is on a secure path to trend to zero. The first order of business is to get the world on a pathway to a CO₂ emissions scenario like RCP4.5 or lower, whereas at present we appear to be headed toward the nightmare of RCP8.5, if not worse. If we fail to avoid that peril, nothing we do about SLCP will matter much.

A multibasket (Daniel et al. 2011) or multitreaty (Jackson 2009) approach is called for, which recognizes the qualitatively different consequences of SLCP emissions and emissions of CO₂ and other gases with millennial impact. The principal challenge, if one is to do something to encourage SLCP mitigation, is to design policy vehicles that provide an incentive for low-cost SLCP mitigation to be done near term, but without allowing it to substitute for CO₂ abatement that would otherwise take place. More importantly, policy vehicles also need to provide an incentive for abatement of SLCP once CO₂ is on track downward but still assure that the SLCP incentives do not result in any increase of cumulative carbon emitted.

DISCLOSURE STATEMENT

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