Atmospheric composition, irreversible climate change, and mitigation policy

S. Solomon[1,2], R. Pierrehumbert[3], D. Matthews[4], J. S. Daniel[5] and P. Friedlingstein[6]

[1]Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139
[2]Department of Atmospheric and Oceanic Sciences, University of Colorado, Boulder, CO 80309
[3]The University of Chicago, Department of the Geophysical Sciences, 5734 S. Ellis Ave, Chicago, IL 60637
[4]Department of Geography, Planning and Environment, Concordia University, Montreal, Quebec, Canada H3G 1M8
[5]Chemical Sciences Division, Earth System Research Laboratory, NOAA, Boulder, CO 80305
[6]College of Engineering, Mathematics and Physical Sciences, University of Exeter, Exeter, EX4 4QF, UK

Abstract

The Earth's atmosphere is changing due to anthropogenic increases of gases and aerosols that influence the planetary energy budget. Policy has long been challenged to ensure that instruments such as the Kyoto Protocol or carbon trading deal with the wide range of lifetimes of these radiative forcing agents. Recent research has sharpened scientific understanding of how climate system time scales interact with the time scales of the forcing agents themselves. This has led to an improved understanding of metrics used to compare different forcing agents, and has prompted consideration of new metrics such as cumulative carbon. Research has also clarified the understanding that short-lived forcing agents can "trim the peak" of coming climate change, while long-lived agents, especially carbon dioxide, will be responsible for at least a millennium of elevated temperatures and altered climate, even if emissions were to cease. We suggest that these vastly differing characteristics imply that a single basket for trading among forcing agents is incompatible with current scientific understanding.

1. Introduction

Anthropogenic increases in the concentrations of greenhouse gases and aerosols perturb the Earth's energy budget, and cause a radiative forcing[1] of the climate system. Collectively, greenhouse gases and aerosols can be considered radiative forcing agents, which lead to either increased (positive forcing) or decreased (negative forcing) global mean temperature, with associated changes in other aspects of climate such as precipitation and sea level rise. Here we briefly survey the range of anthropogenic greenhouse gases and aerosols that contribute to present and future climate change, focusing on time scales of the global anthropogenic climate changes and their implications for mitigation options.

Differences in atmospheric residence times across the suite of anthropogenic forcing agents have long been recognized. As decision makers weigh near-term and long-term mitigation actions and tradeoffs, residence times of forcing agents are important along with social, economic, and political issues, such as climate change impacts, costs, and risks sustained by later versus earlier generations (and how these are valued). Recent research has rekindled and deepened the understanding (advanced by Hansen et al., 1997; Shine et al., 2005) that climate changes caused by anthropogenic increases in gases and aerosols can last considerably longer than the gases or aerosols themselves, due to the key role played by the time scales and processes that govern climate system responses.

[1] Radiative forcing is defined (e.g., IPCC, 2007) as the change in the net irradiance (downward minus upward, generally expressed in W m-2) at the tropopause due to a change in an external driver of the Earth's energy budget, such as, for example, a change in the concentration of carbon dioxide.

55 The climate changes due to the dominant anthropogenic forcing agent, carbon dioxide,
56 should be thought of as essentially irreversible on time scales of at least a thousand years
57 (Matthews and Caldeira, 2008; Plattner et al., 2008; Solomon et al., 2009, 2010).
58
59 The largely irreversible nature of the climate changes due to anthropogenic carbon
60 dioxide has stimulated a great deal of recent research, which is beginning to be
61 considered within the policy community. Some research studies have focused on how
62 cumulative carbon dioxide may represent a new metric of utility for policy, as a result of
63 the identification of a near-linear relationship between its cumulative emissions and
64 resulting global mean warming. In this paper, we discuss the use of cumulative carbon
65 to help frame present and future climate changes and carbon policy formulation. We
66 also briefly summarize several other metrics such as e.g., carbon dioxide equivalent
67 concentration, the global warming potential (GWP) and global temperature change
68 potential (GTP). Finally, we examine how current scientific understanding of the
69 importance of time scales not just of different forcing agents, but also of their interactions
70 with the climate system, sharpens the identification of approaches to formulate effective
71 mitigation policies across a range of radiative forcing agents.
72
73 2. The mix of gases and aerosols contributing to climate change
74
75 A great deal of recent research has focused on understanding changes in atmospheric
76 composition, chemistry, and the individual roles of the range of forcing agents and
77 precursor emissions (leading to the formation of indirect forcing agents after emission) as

78 contributors to observed and future climate change (Forster et al., 2007; Montzka et al.,
79 2011). It is not our goal to review that literature here, but rather to briefly summarize the
80 state of knowledge of contributions of different species to global radiative forcing and
81 time scales of related climate change, and to identify some implications for mitigation
82 policy.

83 The concentrations of the major greenhouse gases carbon dioxide, methane, and nitrous
84 oxide have increased due to human activities, and ice core data show that these gases
85 have now reached concentrations not experienced on Earth in at least several thousand
86 years (Luthi et al., 2008; Joos and Spahni, 2008; MacFarling-Meure et al., 2008).
87 Figure 1 depicts the dramatic increase in carbon dioxide that has taken place over about
88 the past century. The recent rates of increase in CO_2, CH_4, and N_2O are unprecedented
89 in at least 20,000 years (Joos and Spahni, 2008). The abundances of CO_2, N_2O and
90 CH_4 are well-mixed over the globe, and hence their concentration changes (and radiative
91 forcings) are well characterized from data such as that shown in Figure 1; see Table 1.
92

93 If anthropogenic emissions of the various gases were to cease, their concentrations would
94 decline at a rate governed by physical and chemical processes that remove them from the
95 global atmosphere. Most greenhouse gases are destroyed by photochemistry in the
96 Earth's atmosphere, including direct photolysis and attack by highly reactive chemical
97 species such as the OH free radical. Many aerosols are removed largely by washout.
98 Carbon dioxide is a unique greenhouse gas that is subject to a series of removal processes
99 and biogeochemical cycling with the ocean and land biosphere, and even the lithosphere,
100 leading to a very long "tail" characterizing a portion of its removal (Archer et al., 1997).

101 While the carbon dioxide concentration changes and anthropogenic radiative forcing
102 since 1750 are very well established, the relationship of its concentration changes to
103 changes in emission (including those from land use) is much less well characterized, due
104 to the flow of those emissions through the carbon cycle. A few industrial greenhouse
105 gases have lifetimes of many hundreds or even thousands of years, due to their extreme
106 chemical and photochemical stability and represent nearly "immortal" chemicals; in
107 particular, the fully fluorinated compounds such as CF_4, NF_3, and C_2F_6 fall in this
108 category. These gases also are strong absorbers of infrared radiation on a per molecule
109 basis. While these gases are currently present in very small concentrations, like carbon
110 dioxide their contributions to climate change are essentially irreversible on thousand year
111 time scales even if policies were to lead to reduced or zero emissions.

112 Table 1 summarizes the lifetimes (or, in the case of CO_2, multiple removal time scales)
113 that influence the contributions of the range of gases and aerosols to radiative forcing and
114 climate change. Some related uncertainties in lifetimes and distributions are also
115 highlighted.

116 Direct emissions and other human actions (such as land disturbances, and emissions of
117 precursor gases) have increased the atmospheric burdens of particles, including mineral
118 dust, black carbon, sulfate, and organics. Tropospheric ozone has also increased largely
119 as a result of emissions of precursor gases such as nitric oxide and organic molecules
120 including volatiles as well as methane. Indirect forcings linked to atmospheric aerosols
121 involving changes in clouds may also be very important, and are subject to very large
122 uncertainties (Forster et al., 2007). The short atmospheric lifetimes of aerosols and
123 tropospheric ozone lead to very large variations in their abundances depending upon

124 proximity to local sources and transport, increasing the uncertainty in estimates of their
125 global mean forcing as well as its spatial distribution (see Table 1).
126
127 Observations (e.g. of total optical depth by satellites or ground-based methods) constrain
128 the net total optical depth, or the transparency of the atmosphere, and provide information
129 on the total direct radiative forcing due to the sum of all aerosols better than they do the
130 forcing due to individual types of aerosol. Many aerosols are observed to be internal
131 mixtures, i.e., of mixed composition such as sulfate and organics, which substantially
132 affects optical properties and hence radiative forcing (see the review by Kanakidou et al.,
133 2005, and references therein). Aerosols lead to perturbations of the top-of-atmosphere
134 and surface radiation budgets that are highly variable in space, and depend on the place as
135 well as amount of emissions. Limited historical data for emissions or concentrations of
136 aerosols imply far larger uncertainties in their radiative forcings since pre-industrial times
137 than for the well-mixed gases (see Table 1). Current research focuses on understanding
138 the extent to which some regional climate changes may reflect local climate feedbacks to
139 global forcing (e.g., Boer and Yu, 2003a,b), versus local responses to spatially variable
140 forcings. For example, increases in local black carbon and tropospheric ozone (e.g.,
141 Shindell and Faluvegi, 2009) may have contributed to the high rates of warming observed
142 in the Arctic compared to other parts of the globe. Sulfate aerosols (which are present in
143 higher concentrations in the northern hemisphere due to industrial emissions) have been
144 suggested as a driver of changes in the north-south temperature gradients and rainfall
145 patterns (e.g., Rotstayn and Lohmann, 2002; Chang et al., 2011). Shortwave-absorbing
146 aerosols change the vertical distribution of solar absorption, causing energy that would

147 have been absorbed at the surface and communicated upward by convection to be directly
148 absorbed in the atmosphere instead; this can potentially lead to changes in precipitation
149 and atmospheric circulation even in the absence of warming (e.g. Menon et al. 2002). The
150 large uncertainties in the short-lived forcing terms as well as the regional climate signals
151 they may be inducing have heightened interest in their relevance for mitigation policy,
152 and this is discussed further below (see e.g., Ramanthan and Feng, 2008; Jackson, 2009;
153 Hansen et al., 1997; Jacobson, 2002; UNEP, 2011; Shindell et al., 2012).
154
155 3. Metrics

156 Given the very broad diversity of anthropogenic substances with the potential to alter
157 Earth's climate (e.g., CO_2, CH_4, N_2O, SF_6, CFCs, HFC's, absorbing and reflecting
158 aerosols, chemical precursors, etc.), it is a challenging task to compare the climate effect
159 of a unit emission of (for example) carbon dioxide, with one of methane or sulfur
160 dioxide. Nevertheless, there has been a demand for such comparisons, and various
161 metrics have been proposed. The purpose of such metrics is to boil a complex set of
162 influences down to a few numbers that can be used to aid the process of thinking about
163 how different emissions choices would affect future climate. Among other uses, metrics
164 have been used to simplify the formulation of climate-related policy actions, climate-
165 protection treaties and emissions trading schemes. We suggest that to the extent
166 possible, a metric (or set of metrics) should not impose value judgments, least of all
167 hidden value judgments (see Fuglestvedt et al., 2003). Metrics should provide a
168 simplified yet clear set of tools that the policy makers can use to formulate policy
169 implementations to achieve an agreed set of climate protection ends.

3.1 Radiative forcing and CO_2-equivalent concentration

Radiative forcing is one measure of the influence of the burden of a range of forcing agents on the Earth's radiative budget at a given point in time. A closely related metric sometimes used to compare the relative effects of the range of forcing agents is to express them as CO_2-equivalent concentrations, which is the concentration of CO_2 that would cause the same radiative forcing at the chosen time as a given mix of CO_2 and other chemicals (including greenhouse gases and aerosols).

Figure 2 shows the CO_2-equivalent concentration estimates for a range of major forcing agents based on radiative forcing for 2005 from Forster et al. (2007), as given in NRC (2011). The figure shows that among the major forcing agents, by far the largest uncertainties stem from aerosols. Because aerosols represent a substantial negative forcing (cooling effect), this leads to large uncertainty in the net total CO_2-equivalent concentration that is driving current observed global climate change. Current warming represents a transient response that is about half as large as it would become in the long term quasi-equilibrium state if radiative forcing were to be stabilized (NRC, 2011). Therefore, uncertainties in today's total CO_2-equivalent concentration imply large uncertainties in how close current loadings of forcing agents may be to eventually warming the climate by more than the 2°C target noted in the Copenhagen Accord. As Figure 2 shows, uncertainties in aerosols dominate the uncertainties in total net radiative

192 forcing or total CO_2-equivalent concentration. If aerosol forcing is large, then much of
193 the radiative effect of increases in greenhouse gases is currently being masked by
194 cooling, implying a larger climate sensitivity and far greater risk of large future climate
195 change than if aerosol forcing is small.
196
197 A key limitation of radiative forcing or CO_2-equivalent concentrations as metrics is that
198 they do not include any information about the time scale of the impact of the forcing
199 agent. For example, the radiative forcing for a very short-lived forcing agent may be
200 very high at a given time but would drop rapidly if emissions were to decrease, while a
201 longer-lived constituent implies a commitment to further climate change even if
202 emissions were to stop altogether.
203
204 Insofar as short-lived aerosols produce a cooling, their masking of a part of the impact of
205 the large load of long-lived warming agents implies that an unseen long-term
206 commitment has already been made to more future warming (e.g. Armour and Roe, 2011;
207 Ramanathan and Feng, 2008); Hansen describes this as a "Faustian bargain", since short-
208 lived aerosol masking can be accompanied by accumulation of more long-lasting and
209 hence ultimately more dangerous levels of carbon dioxide and other long-lived
210 greenhouse gases in the atmosphere (e.g., Hansen and Lacis, 1990).
211
212 It is evident that other metrics beyond radiative forcing are needed to capture temporal
213 aspects of the climate change problem. One needs to compare not only the effect of
214 various substances on today's climate change but also how current and past emissions

affect future climate change. As will be shown, available metrics all simplify or neglect aspects of temporal information related to individual gases (albeit in different ways), and hence incorporate choices and judgments rather than representing "pure" physical science metrics (Fuglestvedt et al., 2003; Manne and Richels, 2001; O'Neill, 2000; Manning and Reisinger, 2011; Smith and Wigley, 2000; Shine, 2009).

The problem of formulating a metric for comparing climate impacts of emissions of various greenhouse gases is challenging because it requires consideration of the widely differing atmospheric lifetimes of the gases. Emissions metrics are of most interest, since it is emissions (rather than concentrations) that are subject to direct control. The lifetime affects the way concentrations are related to emissions. For a short-lived gas like CH_4, the concentrations are a function of emissions averaged over a relatively short period of time (on the order of a few decades in the case of CH_4). For example, while anthropogenic emissions increase, the CH_4 concentration increases but if anthropogenic emissions of CH_4 were to be kept constant, the concentration of the gas would reach a plateau within a few decades. In contrast, for a very persistent gas like CO_2, the concentration is linked to the cumulative anthropogenic emission since the time when emissions first began; concentrations continue to increase without bound so long as emissions are significantly different from zero. In essence, a fixed reduction of emission rate of a short-lived gas yields a step-reduction in radiative forcing, whereas the same reduction of emission rate of a very long-lived gas only yields a reduction in the rate of growth of radiative forcing.

238 3.2 GWP_h and GTP_h

239

240 The most familiar and widely applied metric for comparing greenhouse gases with
241 disparate atmospheric lifetimes is the Global Warming Potential (GWP). The GWP is
242 defined as the ratio of the time-integrated (over some time horizon) radiative forcing due
243 to a pulse emission of a unit of a given gas, to an emission of the same amount of a
244 reference gas (Forster et al., 2007). This can be expressed as:

245

246 $$GWP_h = \int_!^! \Delta ! \ \Delta C(t)\, dt \ / \int_!^! \Delta ! r \ \Delta C_r(t)\, dt \qquad (1)$$

247

248 where *h* is a specified time horizon, $\Delta C(t)$ is the time series of the change in
249 concentration of the greenhouse gas under consideration (relative to some baseline
250 value), and $\Delta C_r(t)$ that of the reference gas (usually CO_2. as we shall assume throughout
251 the following). ΔA (and ΔA_r) represent the radiative efficiencies due to changes in
252 concentration of the greenhouse gas (and reference gas) following a pulse emission at
253 t=0. In the remainder of this paper, we refer specifically to GWP_h and GTP_h to
254 emphasize the key role of the time horizon. If the pulse is small enough, the radiative
255 forcing is linear relative to the size of the emission pulse; the conventional assumption is
256 therefore that GWP_h is independent of the size of the pulse. This assumption of linearity
257 can lead to substantial errors when the GWP_h is extrapolated from an infinitesimal pulse
258 to very large emissions. Such errors can arise from nonlinearities in the radiative forcing
259 due to changes in concentration of the emitted gas or that of the reference gas CO_2.

260

For gases with short atmospheric lifetimes (e.g. methane), the peak of concentration that immediately follows a pulse in emission decays rapidly to zero, leading to a strong dependence of GWP_h on the timescale over which it is calculated (h in Equation 1). Table 2.14 in Forster et al. (2007) gives GWP_h for a variety of gases, with h = 20, 100 and 500 years. Methane for example, has a 100-year GWP_h (GWP_{100}) of 25, but a GWP_{500} of only 7.6. The choice of time horizon is crudely equivalent to the imposition of a discount rate, albeit a discount rate that varies with lifetime of the gas (Manne and Richels, 2001), and thus represents a value judgment. A choice of small h implies that one should not care that CO_2 saddles the future with an essentially permanent alteration of climate, whereas the choice of a very large h says that one should not care about the transient warming due to short-lived greenhouse gases. Either assumption embeds a judgment regarding whether the near term future is to be valued above the long term future, or vice versa.

An additional concern with the GWP_h is that it represents only the change in integrated forcing due to the emission of different gases, rather than the change in (for example) global-mean temperature. This has led to the proposal of modified metrics, such as the Global Temperature Potential (GTP_h) put forward by Shine et al. (2005). The GTP_h represents the temperature change at some point h in time (rather than time-integrated radiative forcing) resulting from the unit emission of a greenhouse gas, relative to the same emission of carbon dioxide.

In order to illustrate some of the consequences of using GTP_h or GWP_h as climate change metrics for gases of different atmospheric lifetimes, we use a simple two-layer ocean

model to translate radiative forcing and surface temperature change over time. This model is a simpler version of the upwelling-diffusion model used in Shine et al.(2005) to critique GWP_h, and has also been proved useful in analyzing the transient climate response in full general circulation models (Winton et al, 2010; Held et al., 2010). The model consists of a shallow mixed layer with temperature anomaly dT'_{mix} and heat capacity μ_{mix} coupled to a deep ocean with temperature anomaly dT'_{deep} and heat capacity $\mu_{deep} >> \mu_{mix}$. The mixed layer loses heat to space (in part via coupling to the atmosphere) at a rate proportional to its temperature. The equations are

$$m_{mix}\{dT'_{mix}/dt\} = -lT'_{mix} - g(T'_{mix}-T'_{deep}) + \Delta F(t) \qquad (2)$$

$$m_{deep}\{dT'_{deep}/dt\} = -g(T'_{deep}-T'_{mix}) \qquad (3)$$

For constant radiative forcing ΔF, this model[2] has the steady solution $T'_{mix} = T'_{deep} = \Delta F/l$. Hence $1/l$ gives the quasi-equilibrium climate sensitivity. The model relaxes to this equilibrium state on two time scales. On the short time scale (generally a matter of a few years), the mixed layer relaxes to a near-equilibrium with the atmosphere but the deep ocean has not yet had time to warm up, so $T'_{deep} \approx 0$. The transient climate response during this stage is then $T'_{mix} = \Delta F/(l + g)$. 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 time scale T'_{mix} only falls to $T'_{deep}\, g/(l + g)$, and subsequently relaxes to zero on the slow deep ocean time scale. This term is the "recalcitrant warming" due to heat burial in the

[2] The parameters we use in the following are: $\mu_{deep} = 20\mu_{mix} = 200 J/m^2K$ and $\gamma = \lambda = 2 W/m^2K$.

306 deep ocean (Held et al., 2010).

307

308 Figure 3(a) shows the calculated temperature response of the mixed layer in this model
309 due to pulse emissions of greenhouse gases with various lifetimes and forcing
310 efficiencies. In this calculation, the radiative forcing is assumed to be linear in the
311 concentration, and the concentration is assumed to decay exponentially with the stated
312 lifetime. The magnitude of the emission of each gas is chosen so that all correspond to
313 the same value when weighted by GWP_{100}; i.e., for a pulse emission, the radiative forcing
314 integrated over 100 years is identical in all cases. Figure 3 shows that the GWP_{100}
315 weighted emission for a gas with a 10-year (methane-like) lifetime and radiative
316 efficiency can be the same as for the longer lived gases, since a weaker long-term
317 warming can be compensated by a larger short term warming. If the integrated warming
318 over the 100 year period is all we care about, and the damages are linear in warming, then
319 these cases may indeed all be considered to have identical impact in that the methane-like
320 case produces larger damages for a short time, as opposed to a longer period with smaller
321 damages for the longer-lived gases. However, if the objective is to limit the magnitude of
322 warming when the 100 year time span is reached, the use of GWP_{100} greatly exaggerates
323 the importance of the short-lived gas, since virtually all of the warming has disappeared
324 after 100 years. This is a starting point for considering the value of the alternative concept
325 of Global Temperature-Change Potential (GTP_h) as in Shine et al. (2005). Measured in
326 terms of 100-year GTP_h, the 10-year lifetime gas has only 1/4.5 times the impact of e.g.,
327 a 1000 year gas with identical GWP_{100}. The warming after 100 years even in the 10-year
328 lifetime case has not decayed to zero as quickly as the radiative forcing itself (which has

329 decayed by a factor of $4.5x10^{-5}$ over this time). The persistent, or recalcitrant warming
330 arises largely from ocean heat uptake (Solomon et al., 2010). But it should also be
331 emphasized that the 100-year GTP_h does not capture the impact of the large short-term
332 warming from the methane-like case. Such short-term warming could be significant if,
333 for example, the near-term rate of temperature change were leading to adaptation stresses.
334
335 Although GTP_h may be a superior metric to GWP_h for implementing climate protection
336 goals based on a threshold temperature at a given time, it does not resolve the problem of
337 sensitivity to the time frame chosen when computing the metric. Based on 100-year
338 GTP_h, emitting an amount of a 1000-year lifetime gas might be considered to be about
339 twice as bad as an emission of a 50-year lifetime gas; however the long lived gas leads to
340 a warming that is nearly constant over the next 200 years whereas the warming due to the
341 50-year gas has largely disappeared by the end of that time. These two cases result in
342 radically different temperature changes over time and clearly do not represent identical
343 climate outcomes.

344 An additional problem with both GWP_h and GTPh is their dependence on the emission
345 scenario. Figure 3a above represents the case of a pulse emission while Figure 3b shows
346 a second case with constant emissions of a methane-like gas with a 10-year lifetime,
347 compared to constant emissions of a gas with an infinite lifetime (see e.g., Shine, 2005).
348 In both Figures 3a and b, the emissions scenarios were selected such that the GWP_{100}
349 values are equivalent. Emissions are sustained for 200 years, and then set to zero at the
350 year 200. In both cases, the warming continues beyond the point at which the
351 concentration of the gas stabilizes; in the case of the methane-like gas, the concentration

352 (not shown) stabilizes after about 10 years but warming continues to increase, illustrating
353 the continuing warming that occurs despite constant atmospheric concentrations, as the
354 deep ocean takes up heat. For the infinitely long-lived gas, concentrations remain
355 elevated even after emission stops, and warming continues to increase (see next section).
356 Indeed, although both cases are equivalent in terms of GWP_{100}-weighted emissions, the
357 infinite-lifetime case leads to a warming that is not only larger at the end of 200 years,
358 but persists for centuries afterwards. The constant-emissions case illustrates the
359 dependence of GTP_h on the emissions scenario. Neither GWP_h nor GTP_h capture what
360 occurs after emissions cease.

361 As a final example, we have carried out a series of calculations driven by the CO_2 time
362 series computed in Eby et al. (2009). The concentration time series were computed by
363 driving an intermediate-complexity climate-carbon cycle model with historical emissions
364 up to the calendar year 2000, followed by two test scenarios in which the emissions rate
365 rises to a peak after 150 years, and then declines to zero in the subsequent 150 years. The
366 two scenarios shown in Figure 3c show results corresponding to 640GtC and 1280GtC of
367 post-2000 cumulative carbon emissions (see next section). Note that the warming is
368 fairly constant in the 700 years following cessation of emission, given the realistic
369 atmosphere CO_2 used in this case as compared to the infinite-lifetime case shown in
370 Figure 3b. Abating cumulative carbon by 640GtC (the difference between the two
371 emission scenarios shown here) reduces warming by about 0.6K in the two-box model.
372
373 The dashed curves in Figure 3c show what happens if the radiative forcing from CO_2 is
374 augmented by that from methane released at a constant rate between 2000 and 2300, with

375 the total emissions again equivalent to the CO_2 from 640GtC based on weighting with a
376 GWP_{100} of 25 (Forster et al.(2007). The corresponding methane emission rate is 0.31 Gt
377 per year, which is similar to the current anthropogenic emission rate of about 0.35 Gt per
378 year (see http://cdiac.ornl.gov/trends/meth/ch4.htm). Emissions are stopped entirely in
379 2300 in this example. One can think of the curve for 640GtC plus methane (dashed blue
380 line) as the result of deciding to abate CO_2 emissions first and methane later, while the
381 curve with 1280GtC and no methane (solid black line) corresponds to abating methane
382 first and carbon later. It is useful to compare the "Methane First" to that for the "CO_2
383 First" case, recalling that both have the same GWP_{100} weighted emissions. The blue
384 dashed curve ramps up quickly and faster just after 2000 as expected from having more
385 short-lived CH_4. Overall, the two track quite well for the first 100 years (compare the
386 solid black line with the dashed blue line), but thereafter the temperature for "CO_2 First"
387 falls well below that for "Methane First." Moreover, after methane emissions are
388 eliminated, the dashed blue line ("CO_2 First") case converges with the curve for 640GtC
389 alone (solid blue line) within a century, as if methane had never been emitted at all.
390 Figure 3c highlights the comparison between the two curves representing the "Methane
391 First" vs. "CO_2 First" strategies. The shaded region mirrors the analysis of (Daniel et al
392 2011), who used emissions and climate response models that were less idealized. The
393 general lesson to be learned is that over the universe of strategies considered equivalent
394 with regard to GWP_{100}, an emphasis on short-lived forcing agents yields more near-term
395 moderation of warming but comes at the expense of considerably greater long term
396 warming.
397

A comparison of the bottom two curves in Figure 3c, in contrast, illustrates the "peak trimming" benefits of reductions in short-lived forcing agents. However, a comparison of the lower two curves alone gives an incomplete picture of the decision framework. One will always get more warming reduction from doing two beneficial things rather than one beneficial thing, but the real question is whether one would get a still better consequence by putting added resources into further reductions of CO_2 versus applying them to short-lived agents.

From the examples in Figure 3, it is clear that emissions of methane (and similarly other short-lived radiative forcing agents) have a strong bearing on the amount of warming during the time over which they are emitted, but have little lasting consequence for the climate system. By contrast, CO_2 and (and to a lesser extent other long-lived forcing agents) are relevant to both short- and long-term climate warming, and in particular generate warming which persists at significant levels long after emissions are eliminated. These fundamental differences between short- and long-lived radiative forcing agents cannot be captured by either GWP_h or GTP_h metrics, which by design can only provide comparisons for the chosen time horizon. Here we have illustrated key limitations of such an approach over time.

3.3. Irreversibility of CO_2-induced warming, climate commitment, and the cumulative CO_2 emissions metric

As illustrated above, whereas shorter-lived gases and aerosols have a strong bearing on near-future climate changes, warming that persists beyond the 21st century, and particularly warming that persists beyond the period of time that humans emit

422 greenhouses gases, will be primarily determined by how much carbon dioxide is emitted
423 over this period of time. Because of the long lifetime of carbon dioxide in the
424 atmosphere compared to other major greenhouse gases, the long-term warming legacy of
425 anthropogenic greenhouse gases will be primarily determined by CO_2-induced warming.

426 In recent literature, the concept of the irreversibility of climate change due to CO_2
427 emissions was first highlighted by Matthews and Caldeira (2008) based upon results from
428 an Earth Model of Intermediate Complexity (EMIC). This has led to the recognition that
429 cumulative carbon (the total tonnes of carbon emitted) has particular utility for policy.
430 Matthews and Caldeira (2008) showed that if CO_2 emissions were eliminated, globally-
431 averaged temperature stabilized and remained approximately constant for several
432 hundred years; notably, though CO_2 concentrations decreased in the atmosphere,
433 temperatures remained at a nearly constant level, mainly as a result of a declining rate of
434 heat uptake by the ocean that approximately balances the decline in carbon dioxide
435 levels; for a detailed discussion see Solomon et al. (2010). Several other EMIC studies
436 have also demonstrated the irreversibility of CO_2-induced warming. Solomon et al
437 (2009) showed that even after 1000 years of model simulation following the elimination
438 of CO_2 emissions, global temperatures were essentially irreversible, remaining within
439 about half a degree of their peak values for a broad range of emission rates and maximum
440 concentrations. In an intercomparison of eight EMICs, Plattner et al (2008) showed
441 persistence of high global temperatures for at least several centuries following zero
442 emissions across all the models. More comprehensive global climate models require
443 much more computer time and hence have thus far been run for zero emission tests over
444 multiple centuries rather than millennia, and show similar results (Lowe et al 2009 and

445 Gillett et al 2011). These studies have confirmed that irreversibility of CO_2-induced
446 warming is a property of the climate system that is driven by basic properties of the
447 system, notably the carbon and ocean heat timescales, and is not limited to intermediate-
448 complexity models.

449 This body of literature has all contributed to estimating what has been called the "zero-
450 emissions commitment"; that is the anticipated future warming that occurs in the absence
451 of additional future CO_2 emissions. This quantity is distinct from another widely-used
452 definition of committed warming: the "constant-composition commitment," which is
453 defined as the future global temperature change which would be expected under constant
454 concentrations of atmospheric CO_2 (Meehl et al., 2007).

455 The difference between these two measures of committed future warming was
456 highlighted by Matthews and Weaver (2011), and summarized in Figure 4a. Under
457 constant atmospheric CO_2 concentrations, temperatures continue to increase as the
458 climate system slowly adjusts to the current atmospheric forcing from CO_2 in the
459 atmosphere. By contrast, if CO_2 emissions were set to zero, atmospheric CO_2 would
460 decrease over time due to removal by carbon sinks, but global temperature would remain
461 approximately constant for several centuries. This difference can also be seen in the
462 example of the simple model shown above: constant composition of an infinite-lifetime
463 gas after year 200 in Figure 3b leads to increasing global temperatures, whereas zero
464 emissions of CO_2 at the year 2300 in Figure 3c leads to approximately stable global
465 temperatures. Persistent warming over many centuries is especially relevant for
466 understanding impacts including the large sea level rise that occurs in a warmer world
467 due to slow thermal expansion of the deeper parts of the ocean and the potentially very

468 gradual loss of the great ice sheets of Greenland and Antarctica (Meehl et al., 2007 and
469 references therein).

470

471 The difference between the constant-composition and zero-emission commitment can
472 also be understood in terms of the CO_2 emissions associated with each scenario. Figure
473 4b shows the historical emissions in blue associated with both scenarios, and the future
474 emissions in red required to maintain constant CO_2 concentrations at year-2010 levels.
475 Given the required balance between emissions and removal by carbon sinks to maintain
476 constant atmospheric levels, the future emissions associated with a constant-composition
477 scenario are substantially larger than zero; in this example, the total emissions over 300
478 years required to maintain constant atmospheric CO_2 amount to about 250 GtC, or close
479 to half of the total historical CO_2 emissions (about 500 GtC). These future emissions are
480 consistent with the continued future warming associated with constant atmospheric CO_2
481 concentrations. By contrast, zero future emissions is consistent with near-zero additional
482 future warming.

483 As already noted, the removal of anthropogenic CO_2 from the atmosphere involves a
484 multitude of time scales, ranging from a few decades for uptake by the upper ocean and
485 land biosphere, a millennium for uptake by the deep ocean, tens of millennia for
486 carbonate dissolution and weathering to restore ocean alkalinity and allow further uptake,
487 and hundreds of thousands of years for silicate weathering (Archer et al, 1997). The
488 nonlinearity of the carbonate chemistry is important in determining the way climate
489 change relates to larger and larger increases in CO_2. Though the radiative forcing is

490 logarithmic as a function of CO_2 concentration, the carbonate chemistry implies that the
491 fraction of CO_2 that remains in the atmosphere after emission increases with the
492 magnitude of the emission (Eby et al., 2009). Further, the slow decay in radiative forcing
493 due to ocean uptake of carbon following cessation of emissions occurs at roughly the
494 same time scale as the relaxation of the deep ocean temperature towards equilibrium;
495 because these two terms work in opposing directions, the surface temperature attained at
496 the time emissions cease is not only proportional to the cumulative carbon, but is also the
497 temperature which prevails with little change for roughly the next millennium (Matthews
498 and Caldeira, 2008; Solomon et al., 2009; Eby et al., 2009, Solomon et al. 2010).
499
500 The coherence between cumulative emissions of carbon dioxide and global temperature
501 changes has been the subject of several recent studies, and represents a new metric with
502 which to assess the climate response to human CO_2 emissions. Matthews et al (2009) and
503 Allen et al (2009) both identified a strong linear relationship between global temperature
504 change and cumulative carbon emissions. Matthews et al (2009) named this the "carbon-
505 climate response.". In this study, they showed the carbon-climate response is well
506 constrained by both coupled climate-carbon models and historical observations to lie
507 between 1 and 2.1 °C per 1000 GtC emitted (see Figure 5 below, taken from NRC, 2011).
508 Allen et al (2009) used a simpler climate model, but considered a larger range of possible
509 climate sensitivities; as a result, they estimated that the instantaneous temperature change
510 associated with cumulative carbon emissions fell between 1.4 and 2.5 °C per 1000 GtC
511 emitted.

512 Cumulative carbon emissions provides a clear means of estimating the extent of climate
513 warming that will occur from wide range of future CO_2 emissions scenarios.
514 Consequently, the anthropogenic warming that will occur, and which will persist for
515 many subsequent centuries, will be determined to a large extent by the total cumulative
516 emissions which occur between now and the time by which humans stop emitting
517 significant amounts of carbon dioxide. If a tipping point (Lenton et al., 2008) in the
518 earth system were to be experienced at any time in the future, even the immediate
519 cessation of CO2 emissions will be unable to substantially lower the global temperature
520 even on timescales of tens of generations.
521

522 4. Policy Outlook

523

524 Reducing emissions of shorter-lived gases and aerosols (e.g., black carbon) is indeed a
525 highly effective way to reduce climate forcing or the rate of warming on shorter
526 timescales as shown by many authors (see e.g. UNEP, 2011; Jacobson, 2002; Shindell et
527 al., 2012 and references therein), and illustrated here in Figure 6. But Figure 3 above
528 provides key context to better understand choices among policy options. In particular,
529 Figure 3c goes beyond the timescale shown in Figure 6 to illustrate that reductions of
530 short-lived gases or aerosols should be most appropriately thought of as an approach to
531 "trimming the peak" warming (and perhaps the rate of warming) in the near term (but
532 recall the discussion in connection with Figure 3, bearing on the question of choices
533 between efforts put into peak trimming versus additional CO_2 reductions). Furthermore,
534 delays in the abatement of short lived forcing agents imply greater heat storage in the

deep ocean and greater sea level rise; thus, the utility of the peak trimming is affected by when it is implemented as well as by how much. Peak trimming can also reduce the rate of warming, with attendant benefits for the ability of human and natural systems to adapt. Greater benefits in peak trimming are obtained the sooner the emissions are abated (see Held et al., 2010). However, Figure 3c also shows that the long term climate – i.e. the character of the "Anthropocene" – is determined largely by the cumulative carbon emitted. It is noteworthy that the use of GWP_{100} in a policy vehicle would consider the "Methane First" scenario to be equivalent to the "CO_2 First" scenario, but the figure makes clear that the latter yields a far better outcome if one is concerned about the climate changes that last beyond 100 years. Thus Figure 3c demonstrates why trimming the peak cannot substitute for reductions in carbon dioxide emissions that will dominate Earth's climate for many centuries if unabated.

A key policy issue involves the relative reductions to make in the emissions of the range of greenhouse gases. The Kyoto Protocol addressed this issue by placing the regulated greenhouse gases into a single basket and relating their emissions in a common CO2-equivalent emission determined by multiplying actual emissions with the 100-yr GWP_h. Numerous studies have demonstrated that using a single metric in this way has drawbacks arising from the disparity in global lifetimes of the various gases. As we have illustrated here, the choice of a particular time horizon includes value judgments regarding the importance of climate changes at varying times. For example, if a GWP_h with a short time horizon is used in order to better equate short-term climate impacts among gases, the larger relative impact of gases with long lifetimes over long timescales

558 will not be considered. Perhaps more importantly, the use of the GWP_h as the trading
559 metric leads to greenhouse gas trading based on relative integrated radiative forcing,
560 which has a limited connection to temperature change (as shown by the comparison of
561 GTP_h to GWP_h) but probably better represents sea level rise (Smith and Wigley, 2000).
562 Many studies have examined ways to more effectively address near-term and long-term
563 warming (e.g., Manne and Richels, 2001 and others), but the majority of policy
564 discussions have revolved around greenhouse gas metrics for a given time that cannot
565 account for time-varying policy goals.
566
567 The Montreal Protocol regulated ozone-depleting substances (ODSs) that were also
568 characterized by very different lifetimes. This Protocol was highly successful in reducing
569 ozone depletion and took a different approach from that of the Kyoto Protocol. Rather
570 than group all ODSs into a single basket in which production and consumption reductions
571 could be traded using some metric like the ozone depletion potential (ODP), the Montreal
572 Protocol effectively regulated groups of gases (e.g., CFCs, HCFC, halons) and some
573 individual gases (e.g., CH_3CCl_3, CCl_4, CH_3Br) separately. Members of these groups were
574 largely characterized by similar lifetimes. It has been shown that if the Montreal Protocol
575 took an alternative single basket approach, and if trading among ODSs were possible and
576 were performed, the success of the Protocol in limiting short term risks could have been
577 compromised (Daniel et al., 2011).
578
579 The principal conclusion of the discussion presented in this paper is that the scientific
580 basis for trading among all greenhouse gases in one single basket is poor, and a more

581 science-based approach for the Kyoto Protocol (and similar regulatory frameworks)
582 would be to abandon the idea of a single-basket approach altogether. As we have shown,
583 short-lived greenhouse gases or aerosols, and CO_2 are knobs that control quite different
584 aspects of the future climate. It does not appear likely that any single metric will be able
585 to fairly represent both. Yet both time scales are clearly important from the policy
586 viewpoint of risks of different types of future climate changes, such as a possibly slow
587 loss of ice from Greenland and Antarctica over millennia and associated massive sea
588 level rise, versus the potential for rapid increases in the area burned by wildfire in the
589 next decade or two. Thus, the research of the past few years shows even more clearly
590 than previous studies that the existing single-basket GWP_h framework is difficult to
591 justify.
592
593 Many of the problems with GWP_h and GTP_h are not intrinsic to the metrics themselves,
594 but to the imposition of a single time scale when computing the metric. As a minimum, a
595 two-basket approach seems to be needed. One basket could be CO_2, and the metric used
596 to quantify the climate impact of that basket would be cumulative carbon emission
597 (Matthews et al., 2009). Further work is needed to determine whether perfluorocarbons
598 might also be included in this basket through a suitable adjustment of cumulative carbon.
599 The long-term basket should be recognized as the only path to managing long-term risks
600 to the climate. The second basket would include much shorter-lived forcing agents such
601 as CH_4, tropospheric ozone, and black carbon, which could be grouped together and
602 measured by a metric such as the GTP_h. Carbon dioxide can be considered here as well,
603 since its growth is expected to be important for the rate of climate change in the near term

604 (as well as being not only dominant but controlling the changes in the long-term).
605 Reducing short-lived gases or aerosols does nothing to reduce the long-term risk posed
606 by substances such as carbon dioxide. This second basket would explicitly recognize
607 and manage what can be done to reduce warming in the short-term time scale of decades
608 or so, with the choice of time horizon *h* being essential. Such an approach would make
609 explicit that reducing short-lived forcing agents can "trim the peak" of global warming
610 but does not, as is sometimes erroneously stated, "buy time" to deal with carbon and
611 other gases (Biello, 2012), unless one neglects entirely the longer term impacts of current
612 actions. A two-basket framework would require careful and interactive analysis of the
613 science, risks, and value judgments associated with choosing how much and when to
614 reduce the short-lived and long-lived baskets, and we believe that it would result in a
615 clearer path forward for mitigation policy.
616
617
618 **References**

619 Allen, M. R., D. J. Frame, C. Huntingford, C. D. Jones, J. A. Lowe, M. Meinshausen, and
620 N. Meinshausen, Warming caused by cumulative carbon emissions towards the
621 trillionth tonne, Nature, 458 (7242), 1163–1166, doi:10.1038/nature08019, 2009.

622 Archer, D., H. Kheshgi, and E. Maier-Reimer, Multiple timescales for neutralization of
623 fossil fuel CO2. Geophysical Research Letters 24 (4):405-408, 1997.
624
625 Armour, K. C., and G. H. Roe, Climate commitment in an uncertain world, Geophys.
626 Res. Lett., 38, L01707, doi:10.1029/2010GL045850, 2011

627 Biello, D., (2012). http://www.scientificamerican.com/article.cfm?id=how-to-buy-time-
628 to-combat-climate-change-cut-soot-methane.
629 Boer, G.J., and B. Yu, Climate sensitivity and climate state, Clim. Dyn., 21, 167–176,
630 2003a.
631
632 Boer, G.J., and B. Yu, Climate sensitivity and response, Clim. Dyn., 20, 415–429, 2003b.
633
634 Caldeira K., and J. F. Kasting, Insensitivity Of global warming potentials to carbon
635 dioxide emission scenarios, Nature, 366, 251-253. doi:10.1038/366251a0, 1993.
636
637 Chang, C.-Y., J. C. H. Chiang, M. F. Wehner, A. R. Friedman, and R. Ruedy, Sulfate
638 aerosol control of tropical Atlantic climate over the twentieth century, J. Climate,
639 24, 2540–2555. doi: 10.1175/2010JCLI4065.1, 2011.
640
641 Daniel J. S., S. Solomon, T. J. Sanford, M. McFarland, J. S. Fuglestvedt and P.
642 Friedlingstein, Limitations of single-basket trading: lessons from the
643 Montreal Protocol for climate policy, Climatic Change, DOI: 10.1007/s10584-
644 011-0136-3, 2011.
645
646 Eby, M., K. Zickfeld, A. Montenegro, D. Archer, K. J. Meissner, and A. J. Weaver,
647 Lifetime of anthropogenic climate change: millennial time scales of potential CO2
648 and surface temperature perturbations, Journal of Climate 22 (10):2501-2511,
649 DOI: 10.1175/2008JCLI2554.1, 2009.

650

651 Forster, P., et al., Changes in atmospheric constituents and in radiative forcing, in
652 *Climate Change 2007: The Physical Science Basis*, (S. Solomon et al., eds.), pp.
653 129–234, Camb. Univ. Press, 2007.

654

655 Fuglestvedt J. S., T. K. Berntsen, O. Godal, R. Sausen, K. P. Shine and T. Skodvin,
656 Metrics of climate change: assessing radiative forcing and emission indices. Clim
657 Change 58(3):267–331, 2003.

658

659 Gillett, N. P., V. J. Arora, K. Zickfeld, S. J. Marshall, and W. J. Merryfield, Ongoing
660 climate change following a complete cessation of carbon dioxide emissions,
661 Nature Geoscience, 4, 83-87, 2011.

662 Hansen, J.E., and A.A. Lacis, Sun and dust versus greenhouse gases: An assessment of
663 their relative roles in global climate change. *Nature*, **346**, 713-719,
664 doi:10.1038/346713a0, 1990.

665 Hansen, J., M. Sato, and R. Ruedy,Radiative forcing and climate response. J. Geophys.
666 Res. Atmos. 102, 6831–6864. (doi:10.1029/96JD03436), 1997.

667

668 Hansen J et al., Efficacy of climate forcings, J. Geophys. Res., 110, D18104,
669 doi:10.1029/2005JD005776, 2005.

670

671 Held, I. M., M. Winton, K. Takahashi, T. Delworth, F. Zeng, and G. K. Vallis, Probing

672 the Fast and Slow Components of Global Warming by Returning Abruptly to
673 Preindustrial Forcing. J. Climate, 23, 24182427. doi: 10.1175/2009JCLI3466.1,
674 2010.
675
676 Jackson S. C., Parallel pursuit of near-term and long-term climate mitigation. Science
677 326:526–527, 2009.
678 Jacobson, M. Z., Control of fossil-fuel particulate black carbon and organic matter;
679 possibly the most effective method of slowing global warming. J. Geophys. Res.
680 107, 4410–4431. (doi:10.1029/2001JD001376), 2002.
681
682 Joos, F., and R. Spahni, Rates of change in natural and anthropogenic radiative forcing
683 over the past 20000 years, Proc. Nat. Acad. Sci., 105, 1425-1430, doi:
684 10.1073/pnas.0707386105, 2008.
685 Kanakidou, M., J.H. Seinfeld, S.N. Pandis, I. Barnes, F.J. Dentener, M.C. Facchini, R.
686 Van Dingenen, B. Ervens, A. Nenes, C.J. Nielsen, E. Swietlicki, J.P. Putaud, Y.
687 Balkanski, S. Fuzzi, J. Horth, G.K. Moortgat, R. Winterhalter, C.E.L. Myhre, K.
688 Tsigaridis, E. Vignati, E.G. Stephanou, and J. Wilson, Organic aerosol and global
689 climate modelling: A review, Atmos. Chem. Phys, 5, 1053-1123,
690 doi:10.5194/acp-5-1053-2005, 2005.
691 Lenton, T. M., H. Held, E. Kriegler, J. W. Hall, W. Lucht, S. Rahmstorf and H. J.
692 Schellnhuber, Tipping elements in the Earth's climate system. Proc. Nat. Acad.
693 Sci., 105 (6) 1787-1793, 2008.

694 Lowe, J. A., C. Huntingford, S. C. B. Raper, C. D. Jones, S. K. Liddicoat, and L. K.
695 Gohar, How difficult is it to recover from dangerous levels of global warming?,
696 Env. Res. Lett., 4, 014,012, 2009.

697 Luthi, D., et al. 2008 High-resolution carbon dioxide concentration record 650,000–
698 800,000 years before present, Nature 453, 379-382, doi:10.1038/nature06949,
699 2008.

700 MacFarling-Meure, C., D. Etheridge, C. Trudinger, P. Steele, R. Langenfelds, T. van
701 Ommen, A. Smith, and J. Elkins, Law Dome CO2, CH4 and N2O ice core records
702 extended to 2000 years BP, Geophysical Research Letters 33, L14810, 2006.
703 Manne, A. S., and R. G. Richels, An alternative approach to establishing trade-offs
704 among greenhouse gases. Nature 410, 675–677. (doi:10.1038/35070541), 2001.
705
706 Manning, M. and A. Reisinger, Broader perspectives for comparing different greenhouse
707 gases, Phil. Trans. R. Soc. A 369, 1891–1905, doi:10.1098/rsta.2010.0349, 2011.
708
709 Matthews, H. D., and K. Caldeira, Stabilizing climate requires near-zero emissions,
710 Geophys. Res. Lett., 35, L04,705, 2008.

711

712 Matthews, H. D., N. Gillett, P. A. Stott, and K. Zickfeld, The proportionality of global
713 warming to cumulative carbon emissions, Nature, 459, 829–832, 2009.

714 Matthews, H. D. and A. J. Weaver, Committed climate warming, Nature Geoscience 3,
715 142 – 143, doi:10.1038/ngeo813, 2010.

716 Meehl, G. A., et al., Global climate projections, in *Climate Change 2007: The Physical*
717 *Science Basis, Contribution of Working Group I to the Fourth Assessment Report*
718 *of the Intergovernmental Panel on Climate Change* (Solomon, S., et al. Eds.),
719 Camb. Univ. Press, Cambridge, 2007.
720
721 Menon S, J. Hansen, L. Nazarenko, and Y. Luo, Climate effects of black carbon aerosols
722 in China and India, Science 297,2250-2253. DOI: 10.1126/science.1075159,
723 2002.
724
725 Montzka, S.A., E. J. Dlugencky and J. H. Butler, Non-CO2 greenhouse gases and
726 climate change, Nature, 476, 43-50, 2011

727 National Research Council, Climate Stabilization Targets: Emissions, concentrations and
728 impacts over decades to millennia, The National Academies Press, Washington,
729 D.C., 2011.

730 O'Neill, B. C, The jury is still out on global warming potentials. Clim. Change 44, 427–
731 443, doi:10.1023/A:1005582929198, 2000.
732
733 Plattner, G.-K., et al., Long-term climate commitments projected with climate-carbon
734 cycle models, J. Clim., 21, 2721–2751, 2008.
735

736 Ramanathan, V., and Y. Feng, On avoiding dangerous anthropogenic interference with
737 the climate system: Formidable challenges ahead, Proc. Nat. Acad. Sci., 105,
738 14245-14250, doi: 10.1073/pnas.0803838105, 2008.

739 Rotstayn, L. D. and U. Lohmann, Tropical rainfall trends and the indirect aerosol effect,
740 J. Climate ; 15 ; 2103-2116, 2002.

741 Shindell, D. and G. Faluvegi, Climate response to regional radiative forcing during the
742 twentieth century, Nature Geosci., 2, 294-300, doi:10.1038/ngeo473, 2009.
743 Shindell, D., et al., Simultaneously Mitigating Near-Term Climate Change and
744 Improving Human Health and Food Security, Science, 335, 183-189, 2012.
745
746 Shine, K. P., J. S. Fuglestvedt, K. Hailemariam, and N. Stuber, Alternatives to the global
747 warming potential for comparing climate impacts of emissions of greenhouse
748 gases. Clim. Change 68, 281–302. (doi:10.1007/s10584-005-1146-9), 2005.
749
750 Shine, K. P., The global warming potential: the need for an interdisciplinary retrial,
751 Climatic Change, 96, 467-472, doi: 10.1007/s10584-009-9647-6, 2009.
752
753 Shine K.P., T. K. Berntsen, J. S. Fuglestvedt, R. B. S. Skeie, and N. Stuber, Comparing
754 the climate effect of emissions of short- and long-lived climate agents, Phil.
755 Trans. R. Soc. A 365, 1903-1914 doi: 10.1098/rsta.2007.2050, 2007.
756

757 Smith, S. J. and T. M. L. Wigley, Global warming potentials: 1. Climatic implications of
758 emissions reductions. Clim. Change 44, 445–457.

(doi:10.1023/A:1005584914078), 2000.

Solomon, S., G. Kasper Plattner, R. Knutti, and P. Friedlingstein, Irreversible climate change due to carbon dioxide emissions, Proc. Natl. Acad. Sci., 106, 1704–1709, 2009.

Solomon S et al., Persistence of climate changes due to a range of greenhouse gases, Proc. Natl. Acad. Sci., 107,18354-18359, doi: 10.1073/pnas.1006282107, 2010.

UNEP. 2011. Towards an Action Plan for Near-term Climate Protection and Clean Air Benefits, UNEP Science-policy Brief. 17pp.

Winton, M., K. Takahashi, I. M. Held, Importance of ocean heat uptake efficacy to transient climate change, J. Climate, 23, 23332344. doi: 10.1175/2009JCLI3139.1, 2010.

Figure 1 – Carbon dioxide concentrations measured in Antarctic ice cores. The blue curve shows the long record from several cores (available at ftp://ftp.ncdc.noaa.gov/pub/data/paleo/icecore/antarctica/epica_domec/edc-co2-2008.txt), while the red curve and inset shows data for 2000 years prior to 2005 (available at ftp://ftp.ncdc.noaa.gov/pub/data/paleo/icecore/antarctica/law/law2006.txt).

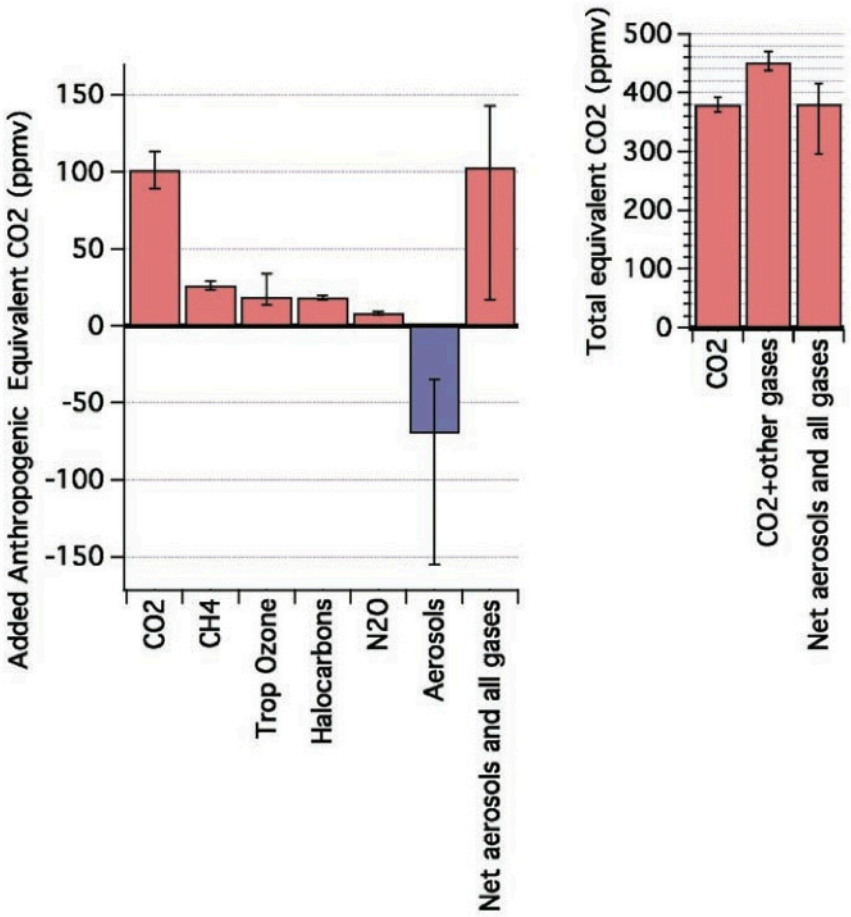

Figure 2 - (left) Best estimates and very likely uncertainty (90% confidence, as in Forster et al., 2007) ranges for aerosols and gas contributions to CO_2-equivalent concentrations for 2005, based on the concentrations of CO_2 that would cause the same radiative forcing as each of these as given in Forster et al. (2007). All major gases contributing more than 0.15 W m–2 are shown. Halocarbons including chlorofluorocarbons, hydrochlorofluorocarbons, hydrofluorocarbons, and perfluorocarbons have been grouped. Direct effects of all aerosols have been grouped together with their indirect effects on clouds. (right) Total CO_2-equivalent concentrations in 2005 for CO_2 only, for CO_2 plus all gases, and for CO_2 plus gases plus aerosols. From Stabilization Targets, NRC, 2011.

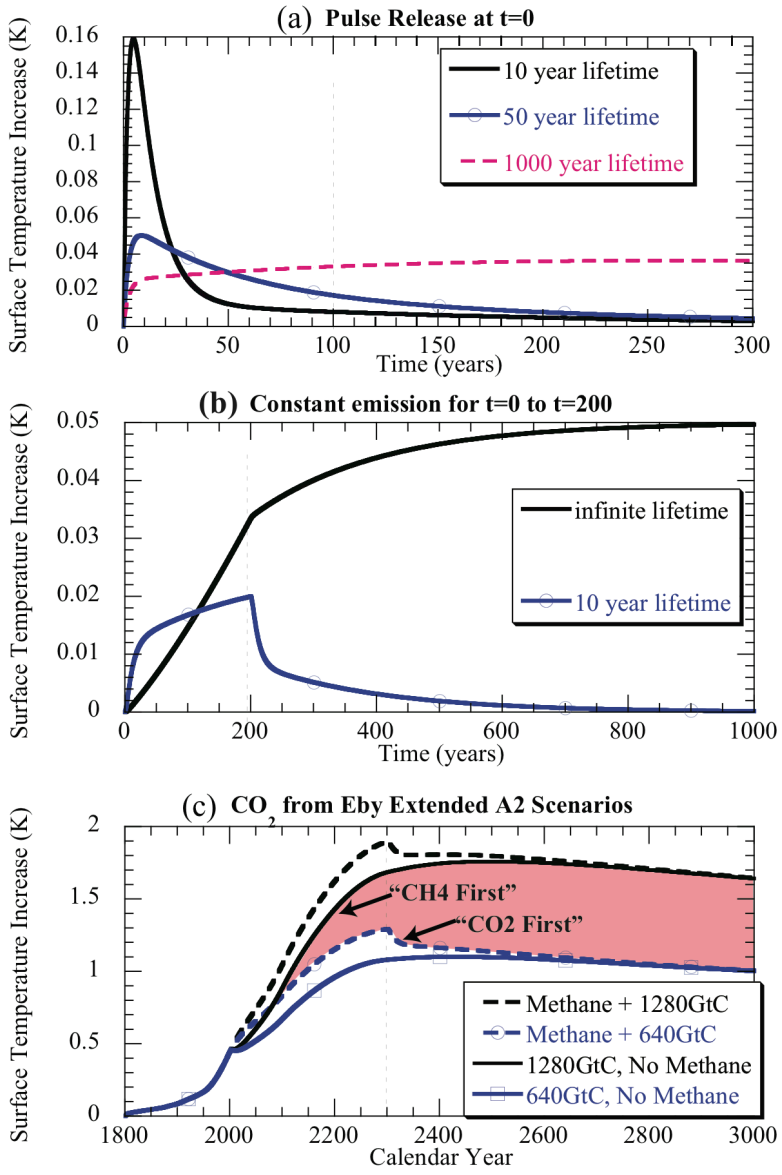

Figure 3 - Surface temperature response of the two-layer ocean model subjected to various time-series of radiative forcing as follows (a) Pulse emission of gases with various lifetimes but identical GWP_{100}. The emission corresponds to an initial radiative forcing of 1 W m^{-2} for the shortest-lived gas. (b) Constant emission rate up to year 200 for an infinite lifetime CO_2-like gas vs. a short-lived methane-like gas having the same GWP_{100}. The total mass of short-lived gas emitted is the same as in the pulse emission calculation shown in (a). (c) Temperature increases from the CO_2 time series in test cases in Eby et al.(2009), corresponding to cumulative carbon emissions of 640 or 1280 GtC between 2000 and 2300, alone or with superposed effect of constant-rate methane emissions with total GWP_{100}-weighted emissions equal to the difference in CO_2 emissions between the two cases; all emissions cease by 2300.

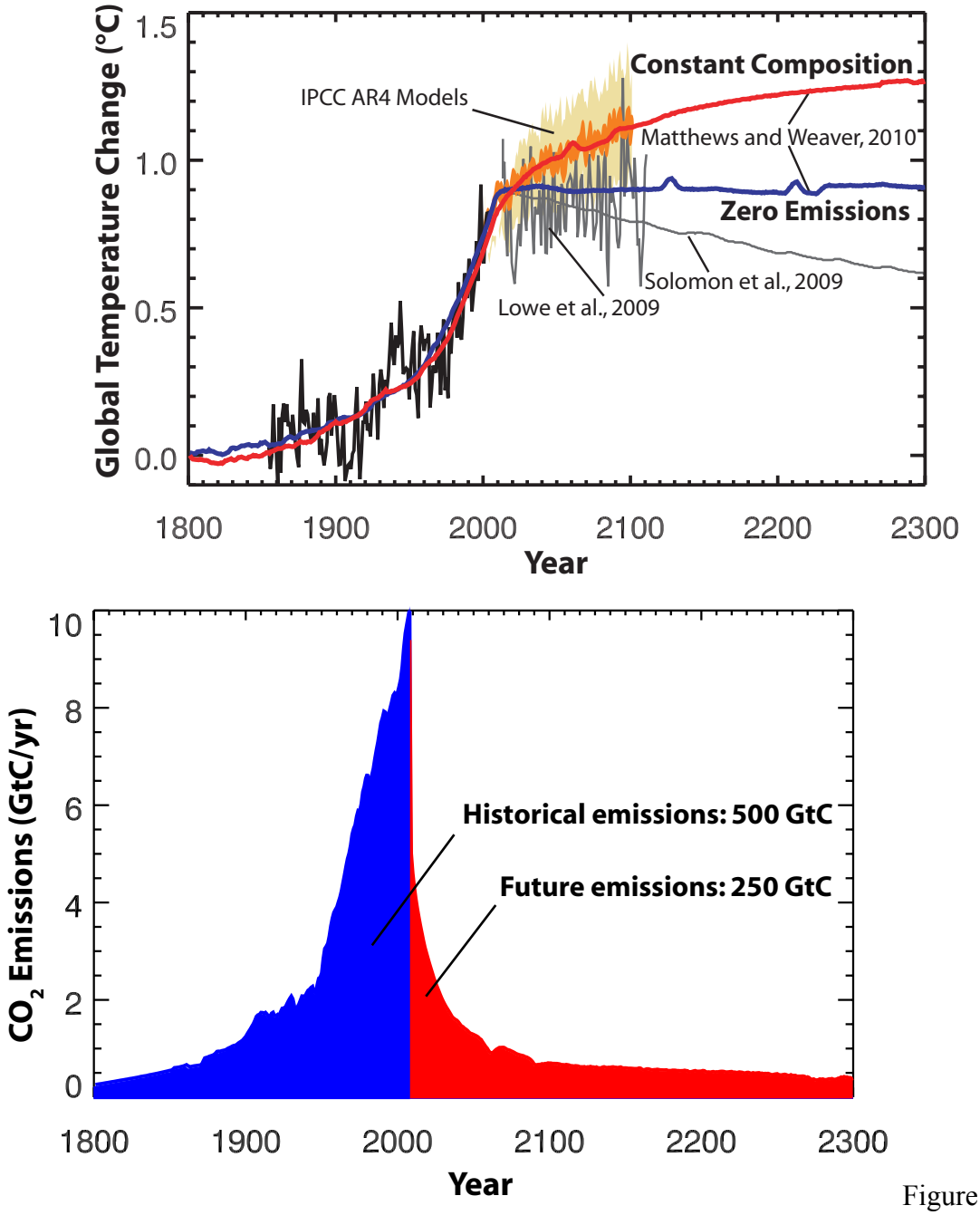

Figure 4 – Figure 4 - Climate response to zero CO_2 emissions, compared to the climate response to constant atmospheric CO_2 concentration. Upper panel shows the global temperature response to zero-emissions from three models (Lowe et al., 2009; Solomon et al., 2009; Matthews and Weaver, 2010) and constant-composition scenarios, as in Matthews and Weaver (2010) and references therein. Lower panel shows the CO_2 emissions scenarios associated with the red and blue lines in panel (a), with cumulative emission given for the historical period (blue shaded area, corresponding to the historical portion of both scenarios) and the future emissions associated with the constant-composition scenario (red shaded area).

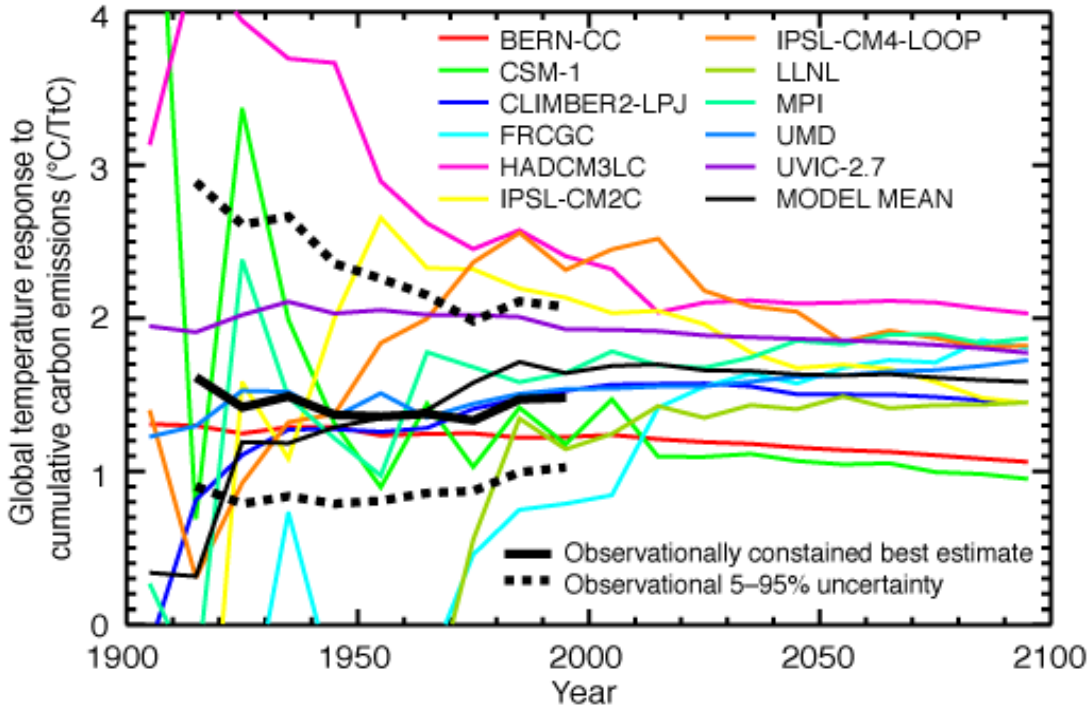

Figure 5 - Climate response to cumulative carbon emissions ("carbon-climate response"), estimated from historical observations of CO_2 emissions and CO_2-attributable temperature changes (thick black line with dashed uncertainty range), as well as from coupled climate-carbon cycle models (colored lines). Both historical observations and model simulations of the 21st century show that the carbon-climate response is approximately constant in time, indicating a linear relationship between cumulative carbon emissions and globally-averaged temperature change. See Matthews et al. (2009) for details.

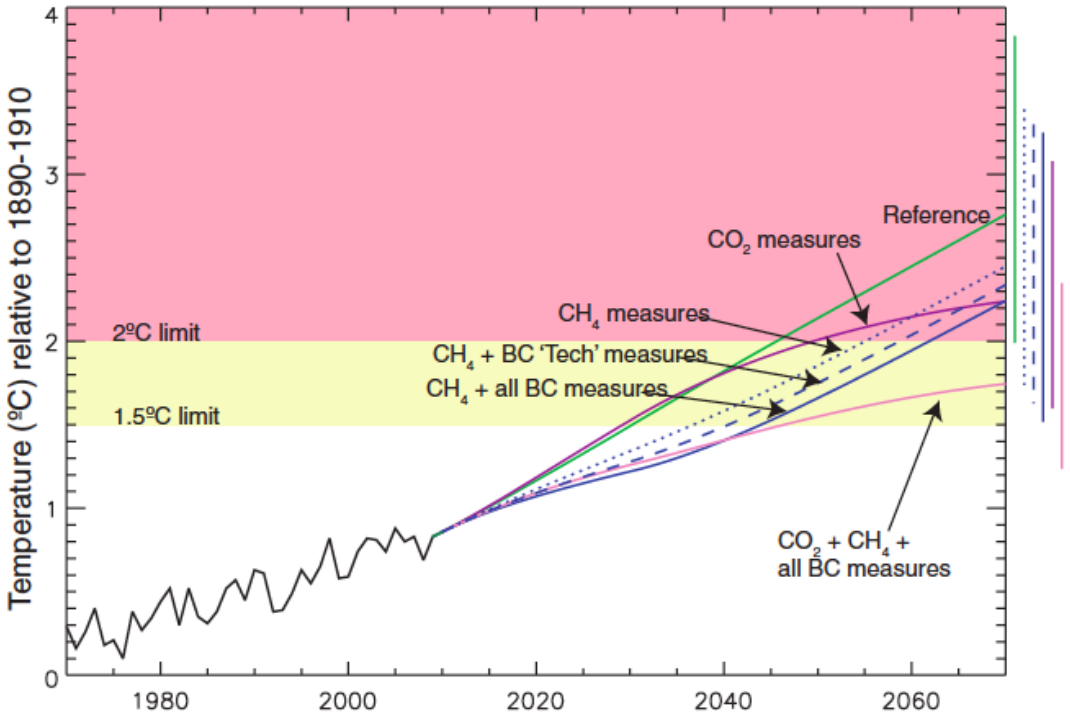

Figure 6 - Observed deviation of temperature to 2009 and projections under various scenarios considered in UNEP (2011) and Shindell et al. (2012); see Shindell et al. (2012) for details. The bulk of the benefits of the assumed CH_4 and black carbon reduction measures are realized by 2040, with the longer term warming being increasingly dependent on carbon dioxide emissions.

Table 1. Atmospheric removals and data required to quantify global radiative forcing for a variety of forcing agents.

Substance	CO_2	Perfluoro-chemicals (CF_4, NF_3, C_2F_6, etc.)	N_2O	Chlorofluoro-carbons ($CFCl_3$, CF_2Cl_2, etc.)	CH_4	Hydrofluoro-carbons (HFC-134a, HCFC-123, etc.)	Tropospheric O_3	Black carbon	Total all aerosols
Atmospheric removal or lifetime	Multiple processes; most removed in 150 years but ≈15-20% remaining for thousands of years	500 to 50000 years, depending on specific gas	≈120 years	≈50 to 1000 years, depending on specific gas	≈10 years	One to two decades to years, depending on specific gas (HFC-23 is exceptional with a lifetime of 270 years).	Weeks	Days	Days
Information on past global changes to quantify radiative forcing	Ice core data for thousands of years; in-situ data for half century quantify global changes well	Some ice core for CF_4. In-situ data quantify current amounts and rates of change well	Ice core data for thousands of years; in-situ data for half century quantify global changes well	Snow (firn) data for hundreds of years; in-situ data for more than three decades quantifies the global changes well	Ice core data for thousands of years; in-situ data for half century quantify global changes well	In-situ data quantifies recent global changes well; clear absence of any significant natural sources avoids need for pre-industrial data	Variable distribution poorly sampled at limited sites; uncertain inferences from satellite data since 1979; very few pre-industrial data.	Extremely variable distribution poorly sampled at limited sites. Some satellite data in last few decades; a few firn data for pre-industrial amounts	Extremely variable distribution poorly sampled at limited sites; some satellite data in last 1-2 decades; no pre-industrial data