Atmospheric composition, irreversible climate change, and mitigation policy

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Abstract

The Earth's atmosphere is changing due to anthropogenic increases of a range 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 the differences between various metrics used to compare emissions of different gases; as a result, there has been an improved understanding of how climate system time scales interact with the time scales of the forcing agents themselves. This has led to consideration of new metrics such as cumulative carbon, and recognition 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. 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 climate changes and their implications for mitigation options.

Differences in atmospheric residence times across the suite of anthropogenic forcing agents have long been recognized. But 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. The climate changes due to the dominant anthropogenic forcing agent, carbon dioxide, should be thought of as essentially irreversible on time scales of at least a thousand years (Matthews and Caldeira, 2008; Plattner et al., 2008; Solomon et al., 2009, 2010).

[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.

52 The largely irreversible nature of the climate changes due to anthropogenic carbon
53 dioxide has stimulated a great deal of recent research, which is beginning to be
54 considered within the policy community. Some research studies have focused on how
55 cumulative carbon dioxide may represent a new metric of utility for policy, as a result of
56 the identification of a near-linear relationship between its cumulative emissions and
57 resulting global mean warming. In this paper, we discuss the use of cumulative carbon
58 to help frame present and future climate changes and policy formulation. We also
59 briefly summarize other metrics such as e.g., carbon dioxide equivalent concentration, the
60 global warming potential (GWP) and global temperature potential (GTP). Finally, we
61 examine how current scientific understanding of the importance of time scales not just of
62 different forcing agents, but also of their interactions with the climate system, sharpens
63 the identification of approaches to formulate effective mitigation policies across a range
64 of radiative forcing agents.
65
66 2. The mix of gases and aerosols contributing to climate change
67
68 A great deal of recent research has focused on understanding changes in atmospheric
69 composition, chemistry, and the individual roles of the range of forcing agents and
70 precursor emissions (leading to the indirect formation of forcing agents after emission) as
71 contributors to observed and future climate change (Forster et al., 2007; Montzka et al.,
72 2011). It is not our goal to review that literature here but rather to briefly summarize the
73 state of knowledge of contributions of different species to global radiative forcing and
74 time scales of related climate change, and to identify some implications for mitigation

75 policy.

76 The concentrations of the major greenhouse gases carbon dioxide, methane, and nitrous
77 oxide have increased due to human activities, and ice core data show that these gases
78 have now reached concentrations not experienced on Earth in many thousands of years
79 (Luthi et al., 2008; Joos and Spahni, 2008; MacFarling-Meure et al., 2008). Figure 1
80 depicts the dramatic increases in trace gases that have taken place over about the past
81 century. The recent rates of increase in CO_2, CH_4, and N_2O are unprecedented in at least
82 20000 years (Joos and Spahni, 2008). The abundances of CO_2, N_2O and CH_4 are well-
83 mixed over the globe, and hence their concentration changes (and radiative forcings) are
84 well characterized from data such as that shown in Figure 1; see also Table 1.
85
86 If anthropogenic emissions of the various gases were to cease, their concentrations would
87 decline at a rate governed by their atmospheric lifetimes or removal processes. Most
88 greenhouse gases are destroyed by photochemical processes in the Earth's atmosphere,
89 including direct photolysis and attack by highly reactive chemical species such as the OH
90 free radical. Many aerosols are removed largely by washout. Carbon dioxide is a unique
91 greenhouse gas that is subject to a series of removal processes and biogeochemical
92 cycling with the ocean and land biosphere, and even the lithosphere. While its
93 concentration changes and anthropogenic radiative forcing since 1750 are very well
94 established, the relationship of concentration changes to anthropogenic emissions is much
95 less well characterized, due to the flow of those emissions through the carbon cycle. A
96 few manmade greenhouse gases have lifetimes of many hundreds or even thousands of
97 years, due to their extreme chemical and photochemical stability and represent nearly

98 'immortal' chemicals; in particular, the fully fluorinated compounds such as CF_4, NF_3,
99 and C_2F_6 fall in this category. These gases also are strong absorbers of infrared radiation
100 on a per molecule basis. While these gases are currently present in very small
101 concentrations, like carbon dioxide their contributions to climate change are essentially
102 irreversible on thousand year time scales even if policies were to lead to reduced or zero
103 emissions.

104 Table 1 summarizes the lifetimes (or, in the case of CO_2, multiple removal time scales)
105 and other important factors that influence the contributions of the range of gases and
106 aerosols to radiative forcing, climate change. Some related uncertainties are also
107 highlighted. Figure 2 summarizes the major contributors to current radiative forcing in
108 terms of CO_2-equivalent concentrations (see below).

109 Direct emissions and other human actions (such as land disturbances, and emissions of
110 precursor gases) have increased the atmospheric burdens of particles, including mineral
111 dust, black carbon, sulfate, and organics. Tropospheric ozone has also increased largely
112 as a result of emissions of precursor gases such as nitric oxide and organics. Indirect
113 forcings linked to atmospheric aerosols involving changes in clouds may also be very
114 important, and are subject to very large uncertainties (Forster et al., 2007). The short
115 atmospheric lifetimes of aerosols and tropospheric ozone lead to very large variations in
116 their abundances depending upon proximity to local sources and transport, increasing the
117 uncertainty in estimates of their global mean forcing (see Table 1).
118
119 Observations (e.g. of total optical depth by satellites or ground-based methods) constrain
120 the net total optical depth, or the transparency of the atmosphere, and provide information

121 on the total direct radiative forcing due to the sum of all aerosols better than they do the
122 forcing due to individual types of aerosols. Many aerosols are observed to be internal
123 mixtures, i.e., of mixed composition such as sulfate and organics, which substantially
124 affects optical properties and hence radiative forcing (see the review by Kanakidou et al.,
125 2005, and references therein). Aerosols lead to perturbations of the top-of-atmosphere
126 and surface radiation budgets that are highly variable in space, and depend on the place as
127 well as amount of emissions. Limited historical data for emissions or concentrations of
128 aerosols imply far larger uncertainties in their radiative forcings since pre-industrial times
129 than for the well-mixed gases (see Table 1). Current research focuses on understanding
130 the extent to which a number of regional climate changes may reflect local climate
131 feedbacks to global forcing (e.g., Boer and Yu, 2003a,b), while others could represent
132 local responses to spatially variable forcings. For example, increases in black carbon and
133 tropospheric ozone (e.g., Shindell and Faluvegi, 2009) may have contributed to the high
134 rates of warming observed in the Arctic compared to other parts of the globe. Sulfate
135 aerosols (which are present in higher concentrations in the northern hemisphere due to
136 industrial emissions) have been suggested as a driver of changes in the north-south
137 temperature gradients and rainfall patterns (e.g., Rotstayn and Lohmann, 2002; Chang et
138 al., 2011). Shortwave-absorbing aerosols change the vertical distribution of solar
139 absorption, causing energy that would have been absorbed at the surface and
140 communicated upward by convection to be directly absorbed in the atmosphere instead;
141 this can potentially lead to changes in precipitation and atmospheric circulation even in
142 the absence of warming (e.g. Menon et al. 2002). The large uncertainties in the short-
143 lived forcing terms as well as the regional climate signals they appear to be inducing have

heightened interest in their relevance for mitigation policy, and this is discussed further below (e.g., Ramanthan and Feng, 2008; Jackson, 2009; Hansen et al., 1997; Jacobson, 2002; UNEP, 2011).

3. Metrics

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

166 3.1 Radiative forcing and CO_2-equivalent concentration

167 Radiative forcing is a metric that has already been introduced and defined above. It is a
168 measure of the influence of the burden of a range of forcing agents on the Earth's
169 radiative budget at a given point in time. A closely related metric sometimes used to
170 compare the relative effects of the range of forcing agents is to express them as CO_2-
171 equivalent concentrations, which is the concentration of CO_2 that would cause the same
172 radiative forcing as a given mix of CO_2 and other chemicals (including greenhouse gases
173 and aerosols).
174
175 Figure 2 shows the CO_2-equivalent concentration estimates for a range of major forcing
176 agents based on radiative forcing for 2005 from Forster et al. (2007), as given in NRC
177 (2011). The figure shows that among the major forcing agents, by far the largest
178 uncertainties stem from aerosols. Because aerosols represent a substantial negative
179 forcing (cooling effect), this leads to large uncertainty in the net total CO_2-equivalent
180 concentration that is driving current observed global climate change. Current warming
181 represents a transient response that is about half as large as it would become in the long
182 term quasi-equilibrium state if radiative forcing were to be stabilized (NRC, 2011).
183 Therefore, uncertainties in today's total CO_2-equivalent concentration imply large
184 uncertainties in how close current loadings of forcing agents are to eventually warming
185 the climate by more than the 2°C target noted in the Copenhagen accord. As Figure 2
186 shows, uncertainties in aerosols dominate the uncertainties in total net radiative forcing or
187 total CO_2-equivalent concentration. If aerosol forcing is large, then much of the
188 radiative effect of increases in greenhouse gases is being masked by cooling, implying a

189 larger climate sensitivity and far greater risk of large future climate change than if aerosol
190 forcing is small.
191
192 A key limitation of radiative forcing or CO_2-equivalent concentrations as metrics is that
193 they do not include information about the time scale of the impact of the forcing agent.
194 The radiative forcing for a very short-lived forcing agent may be very high at a given
195 time but would drop rapidly if emissions were to decrease, while a longer-lived
196 constituent implies a commitment to further climate change even if emissions were to
197 stop altogether.
198
199 Insofar as short-lived aerosols produce a cooling, their masking of a part of the impact of
200 the large load of long-lived warming agents implies that an unseen long-term
201 commitment has already been made to more future warming (e.g. Armour and Roe, 2011;
202 Ramanathan and Feng, 2008); Hansen describes this as a 'Faustian bargain', implying
203 that anthropogenic increases in short-lived aerosol masking implicitly represent a way to
204 allow more long-lasting and hence ultimately more dangerous levels of carbon dioxide
205 and other long-lived greenhouse gases to be accumulated in the atmosphere (e.g., Hansen
206 and Lacis, 1990).
207
208 It is evident that other metrics beyond radiative forcing are needed to capture temporal
209 aspects of the climate change problem. One needs to compare not only the effect of
210 various substances on today's climate change but also how current and past emissions
211 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 metric (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 track emissions averaged over a short period of time (a decade in the case of CH_4. For a very persistent gas, like CO_2, the concentration for all intents and purposes tracks the cumulative 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 one-time reduction in radiative forcing, whereas the same reduction of emission rate of a very long-lived gas yields a reduction of radiative forcing that grows over a long period of time.

3.2 GWP and GTP

The most familiar and widely applied metric for comparing greenhouse gases with disparate atmospheric lifetimes is the Global Warming Potential (GWP). The GWP is

235 defined as the ratio of the time-integrated (over some time horizon) radiative forcing due
236 to a pulse emission of a unit of a given gas, to an emission of the same amount of a
237 reference gas (Forster et al., 2007). This can be expressed as:
238
239 $GWP_h = \int \Delta A \Delta C(t)\, dt \;/\; \int \Delta A_r\, \Delta C_r(t)\, dt$ (1)
240
241 where h is a specified time horizon, $\Delta C(t)$ is the time series of the change in
242 concentration of the greenhouse gas under consideration (relative to some baseline
243 value), and $\Delta C_r(t)$ that of the reference gas (usually CO_2. as we shall assume throughout
244 the following). ΔA (and ΔA_r) represent the radiative efficiencies due to changes in
245 concentration of the greenhouse gas (and reference gas) following a pulse emission at
246 t=0. If the pulse is small enough, the radiative forcing is linear relative to the size of the
247 emission pulse; the conventional assumption is therefore that GWP is independent of the
248 size of the pulse. This assumption of linearity can lead to substantial errors when the
249 GWP is extrapolated from an infinitesimal pulse to very large emissions. Such errors can
250 arise from nonlinearities in the radiative forcing due to changes in concentration of the
251 emitted gas or that of the reference gas CO_2. While the radiative forcing changes due to
252 changes in CO_2 concentrations are not far from linear over a reasonable range of
253 concentrations (Caldeira and Kasting, 1992), this is not generally the case for other gases.
254
255 For gases with short atmospheric lifetimes (e.g. methane), the change in concentration
256 following a pulse emission is dramatic, leading to a strong dependence of GWP on the
257 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 (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 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) put forward by Shine et al. (2005). The GTP represents the temperature change at some point 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 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

281 critique GWP_h, and has also been proved useful in analyzing the transient climate
282 response in full general circulation models (Winton et al, 2010; Held et al., 2010). The
283 model consists of a shallow mixed layer with temperature perturbation dT'_{mix} and heat
284 capacity μ_{mix} coupled to a deep ocean with temperature dT'_{deep} and heat capacity $\mu_{deep} >>$
285 μ_{mix}. The mixed layer loses heat to space (in part via coupling to the atmosphere) at a rate
286 proportional to its temperature. The equations are
287
288 $$\mu_{mix}\{dT'_{mix}/dt\} = -\lambda T'_{mix} - \gamma(T'_{mix}-T'_{deep}) + \Delta F(t) \quad (2)$$
289
290 $$\mu_{deep}\{dT'_{deep}/dt\} = -\gamma(T'_{deep}-T'_{mix}) \quad (3)$$
291
292 For constant radiative forcing ΔF, this model has the steady solution $T'_{mix} = T'_{deep} = \Delta F/\lambda$.[2]
293 Hence $1/\lambda$ gives the quasi-equilibrium climate sensitivity. The model relaxes to this
294 equilibrium state on two time scales. On the short time scale (generally a matter of a few
295 years), the mixed layer relaxes to a near-equilibrium with the atmosphere but the deep
296 ocean has not yet had time to warm up, so $T'_{deep} \approx 0$. The transient climate response
297 during this stage is then $T'_{mix} = \Delta F/(\lambda + \gamma)$. If ΔF is reduced to zero some time after the
298 deep ocean has warmed up to some nonzero value T'_{deep}, then on the short mixed layer
299 time scale T'_{mix} only falls to $T'_{deep}/(\lambda + \gamma)$, and subsequently relaxes to zero on the slow
300 deep ocean time scale. This term is the "recalcitrant warming" due to heat burial in the
301 deep ocean (Held et al., 2010).

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

Figure 3(a) shows the calculated temperature response in this model due to pulse emissions of greenhouse gases with various lifetimes. In this calculation, the radiative forcing is assumed to be linear in the concentration, and the concentration is assumed to decay exponentially with the stated lifetime. The magnitude of the emission of each gas is chosen so that all correspond to the same value when weighted by GWP_{100}; i.e., for a pulse emission, the radiative forcing integrated over 100 years is identical in all cases. Figure 3 shows that the reason the GWP_{100} weighted emission for the gas with a 10-year (methane-like) lifetime is the same as for the longer lived gases is that the weaker long-term warming is compensated by a larger short term warming. If the integrated warming over the 100 year period is all we care about, and the damages are linear in warming, then these cases may indeed all be considered to have identical impact in that the methane-like case produces larger damages for a short time, as opposed to a longer period with smaller damages for the longer-lived gases. However, if the objective is to limit the magnitude of warming when the 100 year time span is reached, use of GWP_{100} greatly exaggerates the importance of the short-lived gas, since virtually all of the warming has disappeared after 100 years. This is a starting point for considering the value of the alternative concept of Global Temperature-Change Potential (GTP) as in Shine et al. (2005). Measured in terms of 100-year GTP, the 10-year lifetime gas has only 1/4.5 times the impact of e.g., a 1000 year gas with identical GWP_{100}. The warming after 100 years even in the 10-year lifetime case has not decayed to zero as quickly as the radiative forcing itself (which has decayed by a factor of 4.5×10^{-5} over this time). The persistent, or recalcitrant warming arises largely from ocean heat uptake (Solomon et al., 2010). But it should also be

325 emphasized that the 100-year GTP does not capture the impact of the large short-term
326 warming from the methane-like case. Such short-term warming could be significant if,
327 for example, the near-term rate of temperature change were leading to adaptation stresses.
328
329 Although GTP_h may be a superior metric to GWP_h for implementing climate protection
330 goals based on a threshold temperature at a given time, it does not resolve the problem of
331 sensitivity to the time frame chosen when computing the metric. Based on 100-year
332 GTP, emitting an amount of a 1000-year lifetime gas might be considered to be about
333 twice as bad as an emission of a 50-year lifetime gas; however the long lived gas leads to
334 a warming that is nearly constant over the next 200 years whereas the warming due to the
335 50-year gas has largely disappeared by the end of that time. These two cases result in
336 radically different temperature changes over time and clearly do not represent identical
337 climate outcomes.

338 An additional problem with both GWP and GTP is their dependence on the emission
339 scenario. Figure 3a above represents the case of a pulse emission while Figure 3b shows
340 a second case with constant emissions of a methane-like gas with a 10-year lifetime,
341 compared to constant emissions of a gas with an infinite lifetime. In both cases, the
342 emissions scenarios were selected such that the GWP_{100} values are equivalent. Emissions
343 are sustained for 200 years, and then set to zero at the year 200. In both cases, the
344 warming continues beyond the point at which the concentration of the gas stabilizes (not
345 shown); in the case of the methane-like gas, the concentration stabilizes after about 10
346 years, and for the infinitely long-lived gas, concentrations stabilize at the point of zero
347 emissions (200 years). The continued warming beyond these points illustrates the

348 committed warming that results under constant atmospheric concentrations (see next
349 section). Though both cases are equivalent in terms of GWP_{100}-weighted emissions, the
350 infinite-lifetime case leads to a warming that is not only larger at the end of 200 years,
351 but persists for centuries afterwards. The constant-emissions case thus illustrates the
352 dependence of GTP on the emissions scenario, and the fact that neither GWP nor GTP
353 capture what occurs after emissions cease.

354 As a final example, we have carried out a series of calculations driven by the CO_2 time
355 series computed in Eby et al. (2009). The concentration time series were computed by
356 driving an intermediate-complexity climate-carbon cycle model with historical emissions
357 up to the calendar year 2000, followed by an extended A2 scenario in which the
358 emissions rate rises to a peak after 150 years, and then declines to zero in the subsequent
359 150 years. The two scenarios shown in Figure 3c show results corresponding to 640GtC
360 and 1280GtC of post-2000 cumulative carbon emissions (see next section). Note that the
361 warming is fairly constant in the 700 years following cessation of emission, given the
362 realistic atmosphere CO_2 used in this case as compared to the infinite-lifetime case shown
363 in Figure 3b. Abating cumulative carbon by 640GtC (the difference between the two
364 cases emission scenarios shown here) reduces warming by about 0.6K in this model.
365
366 The dashed curves in Figure 3c show what happens if the radiative forcing from CO_2 is
367 augmented by that from methane released at a constant rate between 2000 and 2300, with
368 the total emissions again equivalent to the CO_2 from 640GtC based on weighting with a
369 GWP_{100} of 25 (Forster et al.(2007). One can think of the curve for 640GtC plus methane
370 (dashed blue line) as the result of deciding to abate CO_2 emissions first and methane later,

371 while the curve with 1280GtC and no methane (solid black line) corresponds to abating
372 methane first and carbon later. If GWP_{100} were a perfect metric, the temperature for the
373 "Methane First" case would be identical to that for the "CO_2 First" case, insofar as both
374 have the same GWP_{100} weighted emissions. In fact, the two track quite well for the first
375 100 years (compare the solid black line with the dashed blue line), but thereafter the
376 temperature for "CO_2 First" falls well below that for "Methane First." Moreover, after
377 methane emissions are eliminated, the dashed blue line ("CO_2 First") case quickly drops
378 to the curve for 640GtC alone (solid blue line), as if methane had never been emitted at
379 all.
380
381 From this example, it is clear that emissions of methane (and similarly other short-lived
382 radiative forcing agents) have a strong bearing on the amount of warming during the time
383 over which they are emitted, but have little lasting consequences for the climate system.
384 By contrast, CO_2 and (and to a lesser extent other long-lived forcing agents) are relevant
385 to both short- and long-term climate warming, and in particular generate warming which
386 persists long after emissions are eliminated. These fundamental differences between
387 short- and long-lived radiative forcing agents fail to be captured by either GWP or GTP
388 metrics.
389
390 3.3. Irreversibility of CO_2-induced warming, climate commitment, and the cumulative
391 CO_2 emissions metric

392 As illustrated above, whereas shorter-lived gases and aerosols have a strong bearing on
393 near-future climate changes, warming that persists beyond the 21st century, and

394 particularly warming that persists beyond the period of time that humans emit
395 greenhouses gases, will be primarily determined by how much carbon dioxide is emitted
396 over this period of time. Because of the long lifetime of carbon dioxide in the
397 atmosphere compared to other major greenhouse gases, the long-term warming legacy of
398 anthropogenic greenhouse gases will be primarily determined by CO_2-induced warming.

399 In recent literature, the concept of the irreversibility of climate change due to CO_2
400 emissions was first highlighted by Matthews and Caldeira (2008) based upon results from
401 an Earth Model of Intermediate Complexity (EMIC). This has led to the recognition that
402 cumulative carbon (the total tonnes of carbon emitted) has particular utility for policy.
403 Matthews and Caldeira (2008) showed that if CO_2 emissions were eliminated, globally-
404 averaged temperature stabilized and remained approximately constant for several
405 hundred years; notably, though CO_2 concentrations decreased in the atmosphere,
406 temperatures remained at a constant level as a result of a declining rate of heat uptake by
407 the ocean. Several other EMIC studies have also demonstrated the irreversibility of
408 CO_2-induced warming. Solomon et al (2009) showed that even after 1000 years of model
409 simulation following the elimination of CO_2 emissions, global temperatures were
410 essentially irreversible, remaining within about half a degree of their peak values for a
411 broad range of emission rates and maximum concentrations. In an intercomparison of
412 eight EMICs, Plattner et al (2008) showed a consistent persistence of high global
413 temperatures for at least several centuries following zero emissions. More
414 comprehensive global climate models require much more computer time and hence have
415 thus far been run for zero emission tests over multiple centuries rather than millennia, and
416 show similar results (Lowe et al 2009 and Gillett et al 2011). These studies have

417 confirmed that irreversibility of CO_2-induced warming is a property of the climate system
418 which is not limited to intermediate-complexity models.

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

425 The difference between these two measures of committed future warming was
426 highlighted by Matthews and Weaver (2011), and summarized in Figure 4a below.
427 Under constant atmospheric CO_2 concentrations, temperatures continue to increase as the
428 climate system slowly adjusts to the current atmospheric forcing from CO_2 in the
429 atmosphere. By contrast, if CO_2 emissions were set to zero, atmospheric CO_2 would
430 decrease over time due to removal by carbon sinks, but global temperature would remain
431 approximately constant for several centuries. This difference can also be seen in the
432 example of the simple model shown above: constant composition of an infinite-lifetime
433 gas after year 200 in Figure 3b leads to increasing global temperatures, whereas zero
434 emissions of CO_2 at the year 2300 in Figure 3c leads to approximately stable global
435 temperatures.

436 The difference between the constant-composition and zero-emission commitment can
437 also be understood in terms of the CO_2 emissions associated with each scenario. Figure
438 4b shows the historical emissions in blue associated with both scenarios, and the future

emissions in red required to maintain constant CO_2 concentrations at year-2010 levels. Given the required balance between emissions and removal by carbon sinks to maintain constant atmospheric levels, the future emissions associated with a constant-composition scenario are substantially larger than zero; in this example, the total emissions over 300 years required to maintain constant atmospheric CO_2 amount to about 250 GtC, or close to half of the total historical CO_2 emissions (about 500 GtC). These future emissions are consistent with the continued future warming associated with constant atmospheric CO_2 concentrations. By contrast, zero future emissions is consistent with near-zero additional future warming.

As already noted, the removal of anthropogenic CO_2 from the atmosphere involves a multitude of time scales, ranging from a few decades for uptake by the upper ocean, a millennium for uptake by the deep ocean, tens of millennia for carbonate dissolution and weathering to restore ocean alkalinity and allow further uptake, and hundreds of thousands of years for silicate weathering (Archer et al, 1997). The nonlinearity of the carbonate chemistry is important in determining the way climate change relates to net CO_2 emissions. Though the radiative forcing is logarithmic as a function of CO_2 concentration, the carbonate chemistry implies that the fraction of CO_2 that remains in the atmosphere after emission increases with the magnitude of the emission (Eby et al., 2009). Further, the slow decay in radiative forcing due to ocean uptake of carbon following cessation of emissions occurs at roughly the same time scale as the relaxation of the deep ocean temperature towards equilibrium; because these two terms work in opposing directions, the surface temperature attained at the time emissions cease is not only proportional to the cumulative carbon, but is also the temperature which prevails

462 with little change for roughly the next millennium (Matthews and Caldeira, 2008;
463 Solomon et al., 2009; Eby et al., 2009, Solomon et al. 2010].
464
465 The coherence between cumulative emissions of carbon dioxide and global temperature
466 changes has been the subject of several recent studies, and represents a new metric with
467 which to assess the climate response to human CO_2 emissions. Matthews et al (2009) and
468 Allen et al (2009) both identified a strong linear relationship between global temperature
469 change and cumulative carbon emissions. Matthews et al (2009) named this the 'carbon-
470 climate response,' defined as the instantaneous temperature change per unit carbon
471 emitted. In this study, they showed the carbon-climate response is well constrained by
472 both coupled climate-carbon models and historical observations to lie between 1 and 2.1
473 °C per 1000 GtC emitted (see Figure 5 below, taken from NRC, 2011). Allen et al
474 (2009) used a simpler climate model, but considered a larger range of possible climate
475 sensitivities; as a result, they estimated that the instantaneous temperature change
476 associated with cumulative carbon emissions fell between 1.4 and 2.5 °C per 1000 GtC
477 emitted.

478 Cumulative carbon emissions provides a clear means of estimating the extent of climate
479 warming that will occur from wide range of future CO_2 emissions scenarios.
480 Consequently, the magnitude of climate changes which occur in the coming century, and
481 which will persist for many subsequent centuries, will be determined to a large extent by
482 the total cumulative emissions which occur between now and the time by which humans
483 are able to stop emitting carbon dioxide. If a tipping point (Lenton et al., 2008) in the
484 earth system is experienced at some time in the future, even the immediate cessation of

485 CO2 emissions or the emissions of other very long-lived compounds will be unable to
486 substantially lower the global temperature even on timescales of tens of generations.
487 Persistent warming over many centuries is especially relevant for understanding impacts
488 including the large sea level rise that occurs in a warmer world due to slow thermal
489 expansion of the deeper parts of the ocean and the potentially very gradual loss of the
490 great ice sheets of Greenland and Antarctica (Meehl et al., 2007 and references therein).
491

492 4. Policy Outlook

493

494 Reducing emissions of shorter-lived gases and aerosols (e.g., black carbon) is indeed a
495 highly effective way to reduce climate forcing or the rate of warming on shorter
496 timescales (see e.g. UNEP, 2011), as shown by many authors and illustrated here in
497 Figure 6. But Figure 3 above provides key context to better understand choices among
498 policy options. In particular, Figure 3c goes beyond the timescale shown in Figure 6 to
499 illustrate that reductions of short-lived gases or aerosols should be most appropriately
500 thought of an approach to "trimming the peak" warming (and perhaps the rate of
501 warming) in the near term. Furthermore, delays in the abatement of short lived forcing
502 agents imply greater heat storage in the deep ocean; thus, the utility of the peak trimming
503 is strongly affected by when it is implemented. Greater benefits in peak trimming are
504 obtained the sooner the emissions are abated (see Held et al., 2010). However, Figure 3c
505 also shows that the long term climate – i.e. the character of the "Anthropocene" – is
506 determined largely by the cumulative carbon emitted. It is noteworthy that the use of
507 GWP_{100} in a policy vehicle would consider the "Methane First" scenario to be equivalent

508 to the "CO_2 First" scenario, but the figure makes clear that the latter yields a far better
509 outcome if one is concerned about the climate changes that last beyond 100 years. Thus
510 Figure 3c demonstrates why trimming the peak cannot substitute for reductions in carbon
511 dioxide emissions that will dominate Earth's climate for many centuries if unabated.
512
513 A key policy issue involves the relative reductions to make in the emissions of the range
514 of greenhouse gases. The Kyoto Protocol addressed this issue by placing the regulated
515 greenhouse gases into a single basket and relating their emissions in a common CO2-
516 equivalent emission determined by multiplying actual emissions with the 100-yr GWP.
517 Numerous studies have demonstrated that using a single metric in this way has
518 drawbacks arising from the disparity in global lifetimes of the various gases. As we have
519 illustrated here, the choice of a particular time horizon includes value judgments
520 regarding the importance of climate changes at varying times. For example, if a GWP
521 with a short time horizon is used in order to better equate short-term climate impacts
522 among gases, the larger relative impact of gases with long lifetimes over long timescales
523 will not be considered. Perhaps more importantly, the use of the GWP as the trading
524 metric leads to greenhouse gas trading based on relative integrated radiative forcing,
525 which has a limited connection to temperature change (as shown by the comparison of
526 GTP to GWP) but probably better represents sea level rise (Smith and Wigley, 2000).
527 Many studies have examined ways to more effectively address near-term and long-term
528 warming (e.g., Manne and Richels, 2001 and others), but the majority of policy
529 discussions have revolved around fixed greenhouse gas metrics that cannot account for
530 time-varying policy goals.

The Montreal Protocol regulated ozone-depleting substances (ODSs) that were also characterized by very different lifetimes. This Protocol was highly successful in reducing ozone depletion and took a different approach from that of the Kyoto Protocol. Rather than group all ODSs into a single basket in which production and consumption reductions could be traded using some metric like the ozone depletion potential (ODP), the Montreal Protocol effectively regulated groups of gases (e.g., CFCs, HCFC, halons) and some individual gases (e.g., CH_3CCl_3, CCl_4, CH_3Br) separately. Members of these groups were largely characterized by similar lifetimes. It has been shown that if the Montreal Protocol took an alternative single basket approach, and if trading among ODSs were possible and were performed, the success of the Protocol in limiting short term risks could have been compromised (Daniel et al., 2011).

The principal conclusion of the discussion presented in this paper is that the scientific basis for trading among all greenhouse gases in one single basket is poor, and a more defensible approach for the Kyoto Protocol (and similar regulatory frameworks) would be to abandon the idea of a single-basket approach altogether. As we have shown, short-lived greenhouse gases or aerosols, and CO_2 are knobs that control quite different aspects of the future climate. It does not appear likely that any metric will be able to fairly represent both. Yet both time scales are clearly important from the policy viewpoint of risks of different types of future climate changes, such as a possibly slow loss of ice from Greenland and Antarctica over millennia and associated massive sea level rise, versus the potential for rapid increases in the area burned by wildfire in the next decade or two.

554 Thus, the research of the past few years shows even more clearly than previous studies
555 that the existing single-basket GWP framework is difficult to justify.
556
557 Many of the problems with GWP and GTP are not intrinsic to the metrics themselves, but
558 to the imposition of a single time scale when computing the metric. As a minimum, a
559 two-basket approach seems to be needed. One basket could be CO_2, and the metric used
560 to quantify the climate impact of that basket would be cumulative carbon emission
561 (Matthews et al., 2009). Further work is needed to determine whether perfluorocarbons
562 might also be included in this basked through a suitable adjustment of cumulative carbon.
563 The other basket would include much shorter-lived forcing agents such as CH_4,
564 tropospheric ozone, and black carbon, which could be grouped together and measured by
565 a metric such as the GTP. This would require careful and interactive analysis of the
566 science, risks, and value judgments associated with choosing how much and when to
567 reduce the short-lived and long-lived baskets, and we believe that it would result in a
568 clearer path forward for mitigation policy.
569
570
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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	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

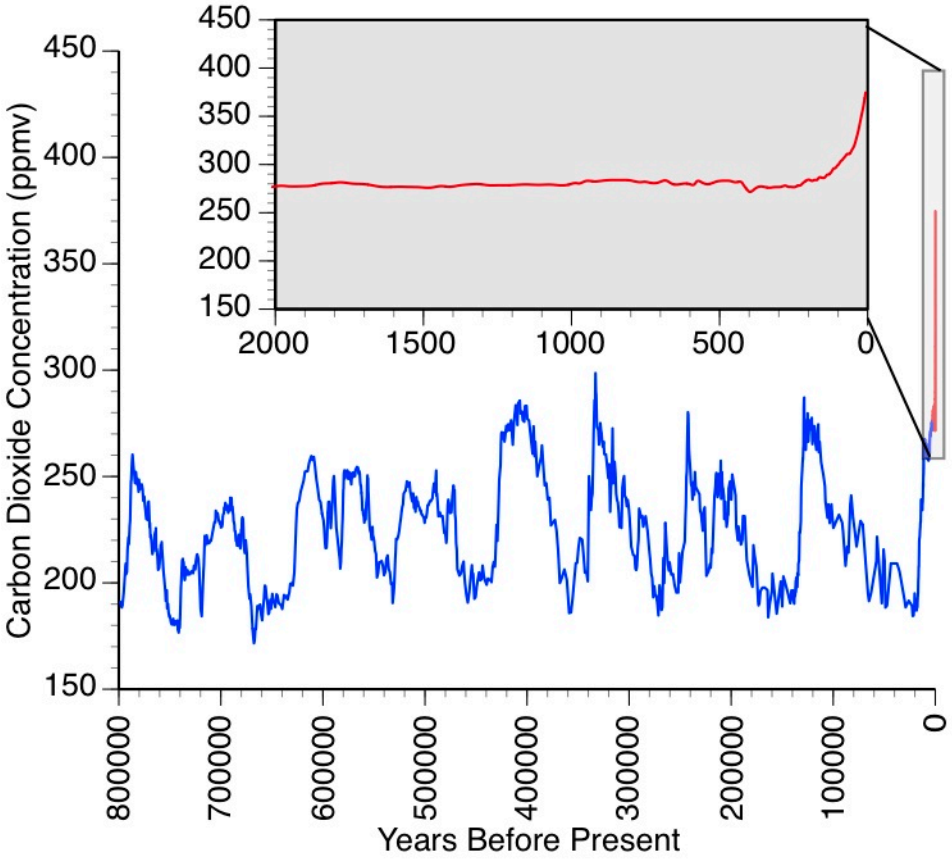

Figure 1 – Carbon dioxide concentrations measured in Antarctic ice cores. The blue curve shows the long record from several cores (available at ftp://ftp.ncdc.noaa.gov/pub/data/paleo/icecore/antarctica/epica_domec/edc-co2-2008.txt), while the red curve and inset shows data for 2000 years prior to 2005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

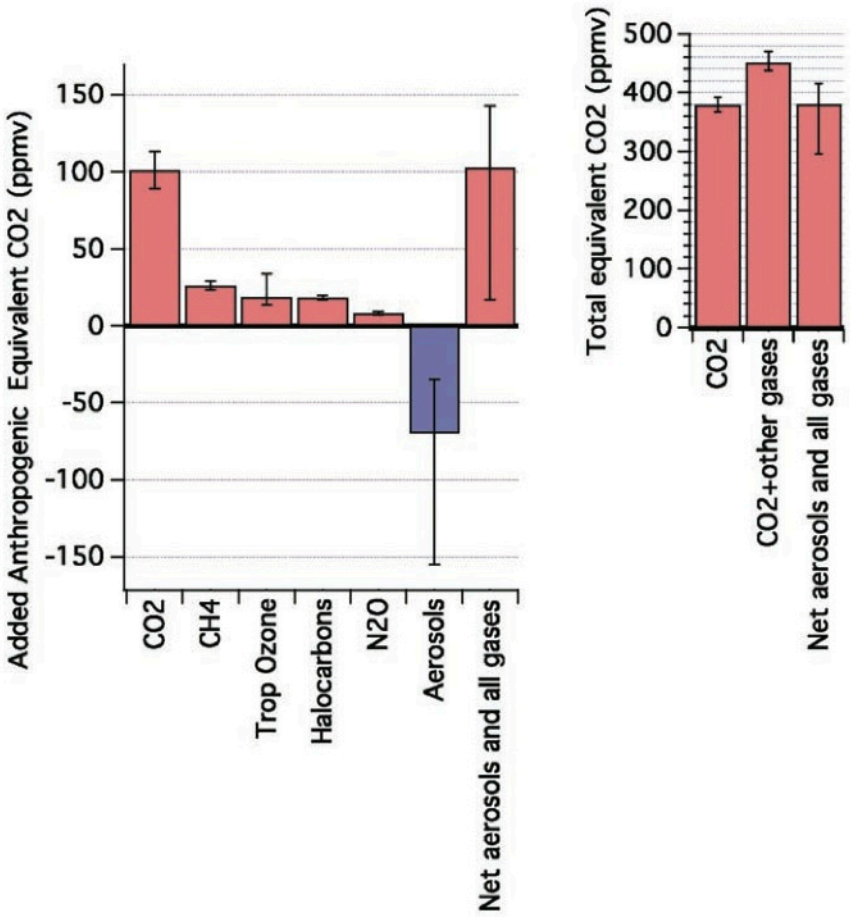

Figure 2 - (left) Best estimates and very likely uncertainty ranges for aerosols and gas contributions to CO_2-equivalent concentrations for 2005, based on the radiative forcing given in Forster et al. (2007). All major gases contributing more than 0.15 W m–2 are shown. Halocarbons including chlorofluorocarbons, 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.

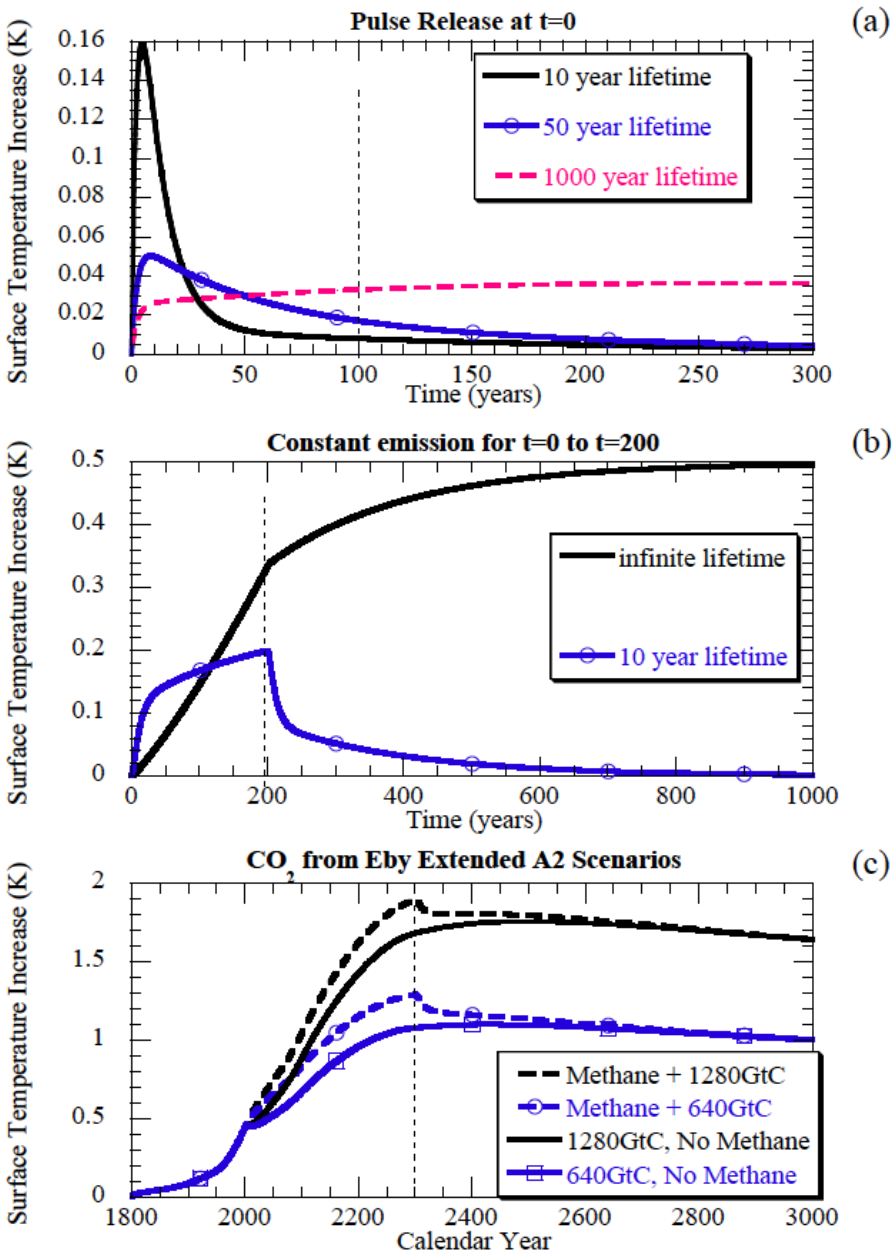

Figure 3 - Surface temperature response of the two-layer ocean model subjected to various time-series of radiative forcing as follows (a) Pulse emission of gases with various lifetimes but identical GWP_{100}. (b) Constant emission rate up to year 200 for an infinite lifetime CO_2-like gas vs. a short-lived methane-like gas having the same GWP_{100}. (c) CO_2 time series from the extended A2 scenarios in Eby et al.(2009), corresponding to cumulative carbon emissions of 640 or 1280 GtC after year 2000, 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.

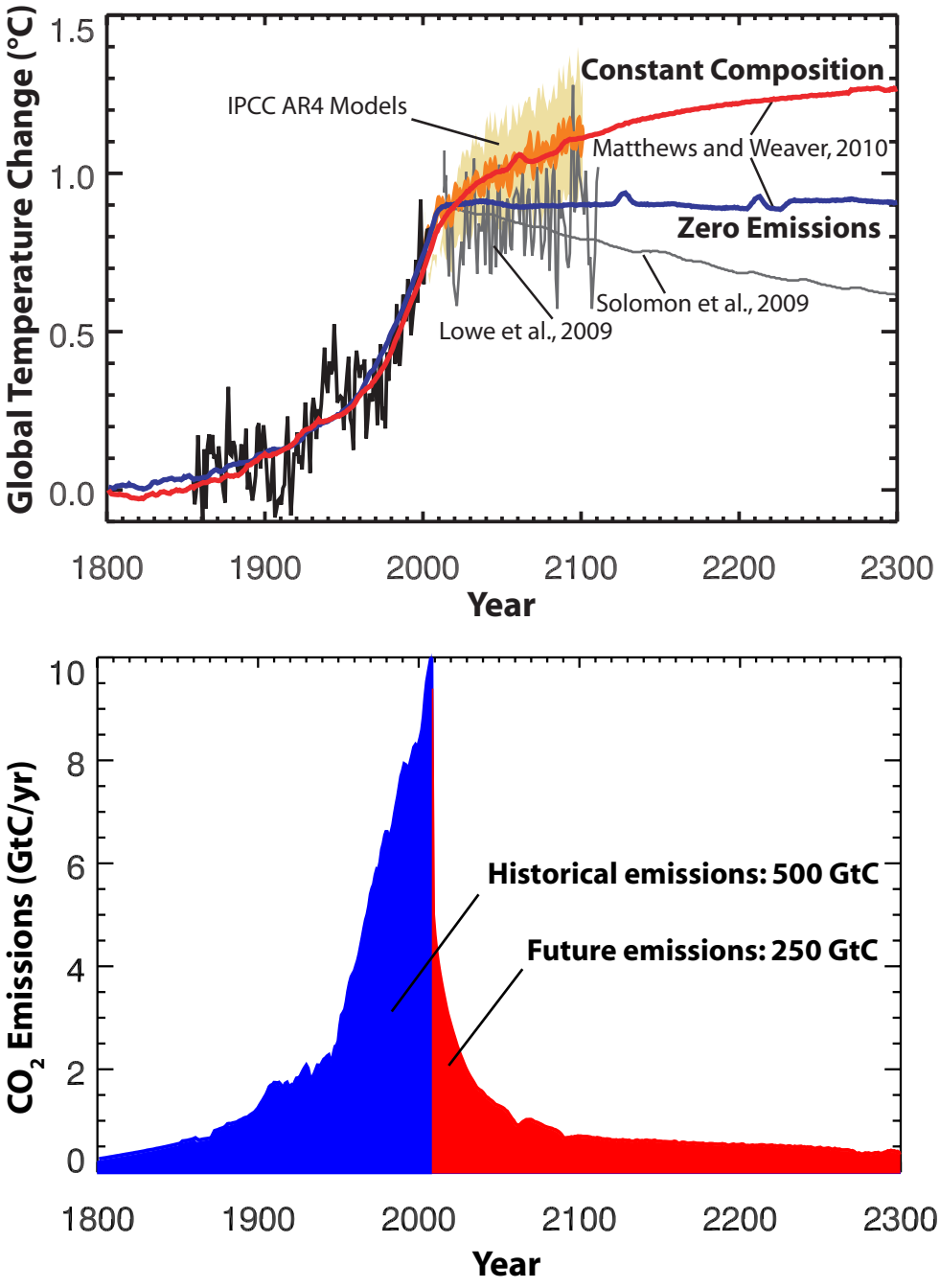

Figure 4 - Climate response to zero CO_2 emissions, compared to the climate response to constant atmospheric CO_2 concentration. Panel (a) shows the global temperature response to zero-emissions and constant-composition scenarios, as in Matthews and Weaver (2010). Panel (b) shows the CO_2 emissions scenarios 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).

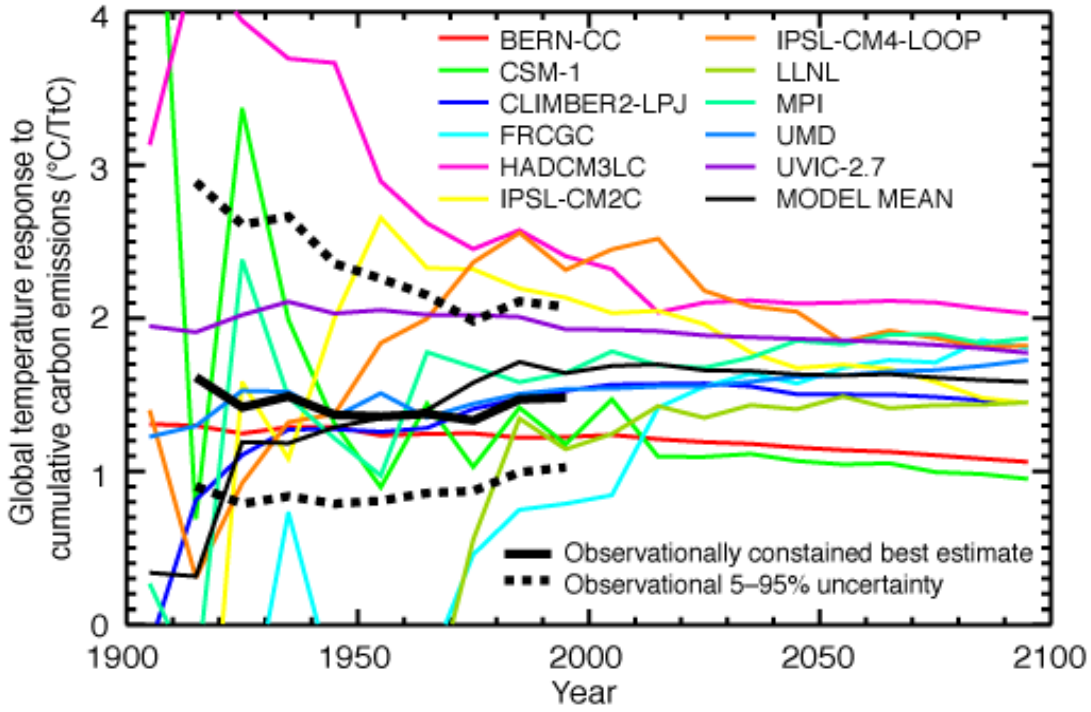

Figure 5 - Climate response to cumulative carbon emissions ("carbon-climate response"), estimated from historical observations of CO_2 emissions and CO_2-attributable temperature changes (thick black line with dashed uncertainty range), as well as from coupled climate-carbon cycle models (colored lines). 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.

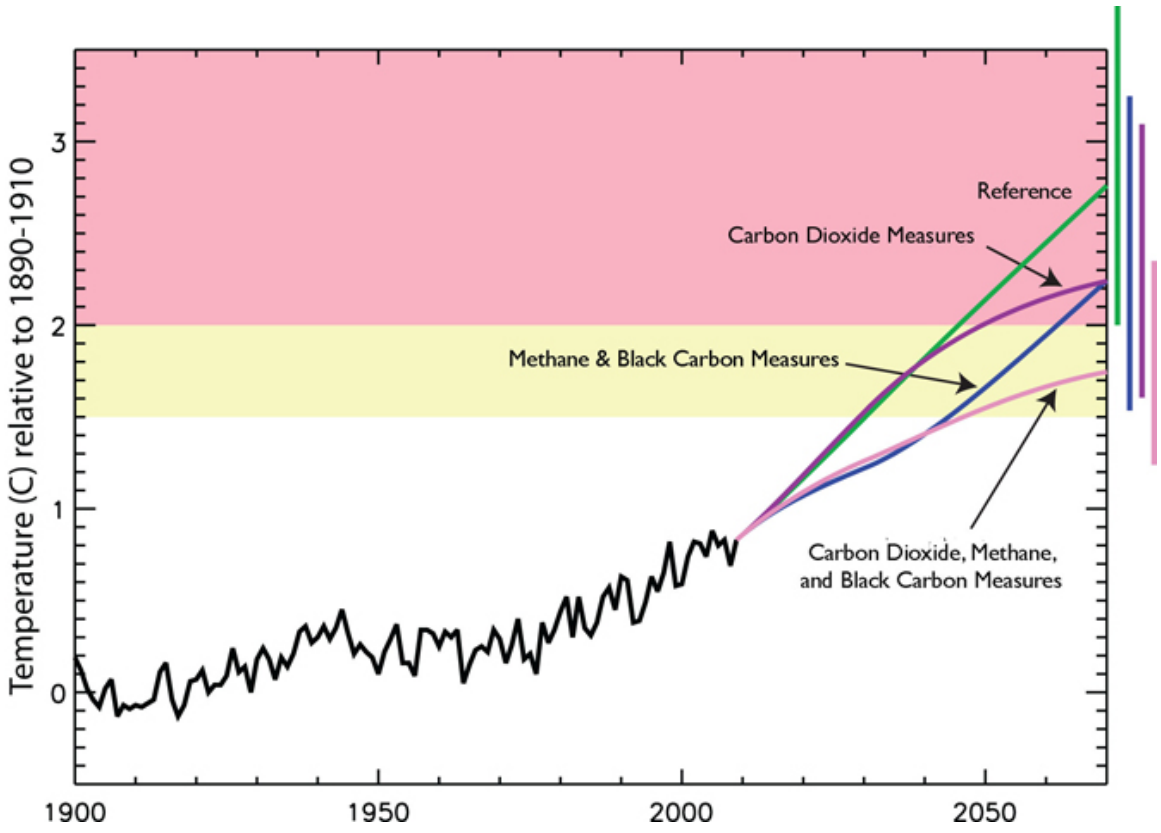

Figure 6 - Observed deviation of temperature to 2009 and projections under various scenarios considered in UNEP (2011). The bulk of the benefits of the assumed CH_4 and black carbon reduction measures are realized by 2040, with the longer term warming being increasingly dependent on carbon dioxide emissions.