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Managing short-lived climate forcers in curbing climate change: an atmospheric chemistry synopsis

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Abstract The Montreal Protocol has set an extraordinary example by applying scientific discoveries, technological innovations, and swift political actions to solving one of the most urgent environmental problems facing humans. With its ongoing implementation, the stratospheric ozone is expected to return to its 1980 levels around mid-twenty-first century. In addition, the Montreal Protocol has contributed to mitigating climate change by reducing the emissions of certain greenhouse gases. The management of several short-lived climate forcers, including hydrofluorocarbons, tropospheric ozone, black carbon, and methane, is worthy of consideration as a fast-response, near-term measure to curb climate change, while international treaties to reduce the emissions of long-lived climate forcers, such as carbon dioxide, are under discussion. This paper aims to provide a concise overview of the scientific concepts and atmospheric processes behind these policy considerations. The focus is on the fundamental atmospheric chemistry that provides the basis for a co-benefits approach in mitigating both climate change and stratospheric ozone depletion.

Keywords Short-lived climate forcers · Hydrofluorocarbons · Montreal protocol · Stratospheric ozone depletion · Greenhouse gases · Climate change

The globally averaged concentrations (in mole fractions) of the three key greenhouse gases (GHGs), carbon dioxide (CO₂), methane (CH₄), and nitrous oxide (N₂O), reached new highs in 2013, with CO₂ at 396.0±0.1 ppm, CH₄ at 1824±2 ppb, and N₂O at 325.9±0.1 ppb, according to the

latest measurements by the World Meteorological Organization (WMO 2014a). These values are, respectively, 42, 153, and 21 % above the pre-industrial (before 1750) levels, reflecting large anthropogenic contributions. Of particular alarm, the atmospheric increase of CO₂ from 2012 to 2013 was 2.9 ppm, the largest year-to-year change in the last 30 years. There is now overwhelming evidence that CO₂ and other climate forcers from human activities have caused the globally averaged surface temperature increase of almost 1 °C from 1880 to 2012 (commonly dubbed “global warming”) and other climate-related changes in the atmosphere, oceans, lands, and ecosystems (IPCC 2013). Many of the observed changes are unprecedented over decades to millennia. Taking immediate and effective measures to curb climate change has become one of the most challenging and critical issues facing humans.

In contrast to the enormous success of the Montreal Protocol in mitigating stratospheric ozone depletion by eliminating ozone-depleting substances (ODSs), efforts to reach an international treaty to curb climate change by limiting major GHGs have yet to come to fruition. For example, the 1997 Kyoto Protocol targeted GHGs but was only for developed countries, had a short duration, and was not legally binding. CO₂, responsible for about 84 % of the increase in radiative forcing¹ in the past decade (WMO 2014a), is a long-lived climate forcer (LLCF). The long lifetime of CO₂ in the atmosphere (centuries to millennia) means that its concentration will continue to rise and remain high for centuries or longer even if CO₂ emissions were to be cut today. In order to avoid the globally averaged surface temperature from rising 2 °C above the preindustrial level, widely thought to be the “tipping

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¹ Radiative forcing: a measure of how a climate forcing agent influences Earth’s energy balance, with a positive value indicating a net heat gain to the lower atmosphere, which leads to a globally averaged surface temperature increase, and a negative value indicating a net heat loss.

point” for the Earth’s climate system, there is an urgent need to identify fast-action measures that can be implemented within 5–10 years and produce a climate response within decades (Molina et al. 2009; Meinshausen et al. 2009).

Recently, it has been proposed by scientists and policy scholars that emissions of short-lived climate forcers or pollutants (SLCFs or SLCPs), including hydrofluorocarbons (HFCs), tropospheric ozone (O₃), black carbon (BC), and CH₄, should be limited to help mitigate climate change in the next few decades (Shoemaker et al. 2013; Burney et al. 2013; Molina et al. 2009). Modeling studies indicate that the combined mitigation of SLCFs can avoid global warming by as much as 0.6 °C by mid-twenty-first century, with about 0.1 °C of this avoided warming attributed to the mitigation of HFCs (Ramanathan and Xu 2010; Shindell et al. 2012; Xu et al. 2013). In the policy arena, strides are already being made. For example, the US Department of State announced in 2012 the “Climate and Clean Air Coalition to Reduce Short-Lived Climate Pollutants” (CCAC), a global initiative aiming to bring about benefits for climate, health, food, and energy simultaneously, with partners including over 30 countries and over 50 non-state partners such as the WMO, the World Bank, the European Commission, and the United Nations Environment Programme. It has been established that many ODSs are also potent GHGs (see, for example, Table 1). Their concentrations remain considerable in the atmosphere, and there are new emissions from the “banks” of ODSs (such as HFCs) contained in operating and abandoned refrigeration, air conditioning, foam, and fire protection systems (Andersen and Sarma 2002; Velders et al. 2014). By the latest estimate, a fast phase-down of HFCs by 2020 would prevent up to 210 billion tonnes (Gt) of CO₂-equivalent (CO₂-eq) emissions by 2050² and avoid up to about 0.5 °C of additional warming by 2100 (Xu et al. 2013; Velders et al. 2014). While measures like these should not delay, but rather complement, the emission cuts of LLCFs (Smith and Mizrahi 2013), they appear promising in short-term effects to curb the warming (Velders et al. 2009, 2012; Shoemaker et al. 2013). In particular, with the expected increase in the use of HFCs to replace chlorofluorocarbons (CFCs) and hydrochlorofluorocarbons (HCFCs) worldwide (IPCC/TEAP 2005; TEAP 2010; Velders et al. 2012), the call for the inclusion of HFCs under a further amendment to the Montreal Protocol, which only limits CFCs and HCFCs as of now, is on the rise (Velders et al. 2012, 2014; Molina et al. 2009; Shoemaker et al. 2013; Andersen et al. 2013; Zaelke and Borgford-Parnell 2014, in this issue).

² Velders GJM, Solomon S and Daniel JS (2014) Growth of climate change commitments from HFC banks and emissions. *Atmos. Chem. Phys.* 14(9):4563–4572 (“If, for example, HFC production were to be phased out in 2020 instead of 2050, not only could about 91–146 Gt CO₂-eq of cumulative emission be avoided from 2020 to 2050, but an additional bank of about 39–64 Gt CO₂-eq could also be avoided in 2050.” The total ranges from 130 to 210 Gt CO₂-eq.)

This paper seeks to provide a concise overview of the relevant scientific concepts, processes, and the interplay of multiple components in the atmosphere, mainly from an atmospheric chemistry perspective. The readers can refer to some classical references in the field of atmospheric chemistry and physics for more in-depth discussions (Seinfeld and Pandis 2006; Spiro et al. 2012). This paper aims to help readers gain a quick understanding of the basic science relevant for the policy discussions in this special issue:

1. Layers of the atmosphere
2. The paradox of ozone: stratospheric depletion vs. tropospheric warming
3. CFCs, HCFCs, and HFCs: ozone depleting potential (ODP) and global warming potential (GWP)
4. Other short-lived climate forcers
5. Interplay of climate change and ozone depletion: making science-based policies

Layers of the atmosphere

The Earth’s atmosphere can be divided, based on the temperature variation with altitude, into several layers, as indicated in Fig. 1. In the lower atmosphere, generally considered to extend up to about 50 km in altitude, two layers—the troposphere and the stratosphere—are most relevant to our discussions.

The troposphere is the lowest layer, extending from the Earth’s surface to the tropopause, the height of which varies with latitude and time of year but is typically ~10 to 15 km in altitude. In this layer, temperature decreases with altitude; as a result, there is rapid mixing of air. Most weather events, as well as air pollution, occur in this layer. The stratosphere extends from the tropopause to the stratopause, which is typically ~45 to 55 km in altitude. In this layer, temperature increases with altitude due to the absorption of UV solar radiation by the stratospheric ozone layer. As a result, the mixing of air is slow in this rather quiescent layer. This has significant consequences for the various chemicals (including ODSs) that reach the stratosphere, such as CFCs, HCFCs, and HFCs. They reside in the stratosphere for a considerably long time.

The paradox of ozone: stratospheric depletion and tropospheric warming

The US Environmental Protection Agency (EPA) characterized in layman’s terms ozone’s roles as “good up high and bad nearby.” Ozone’s seemingly paradoxical roles relate to its vertical distribution in the atmosphere: ozone

Table 1 Atmospheric lifetime, 100-year global warming potential (GWP) and ozone depletion potential (ODP) for selected CFCs, halons, HCFCs, and HFCs

Chemical species	Chemical formula	100-year GWP	ODP	Atmospheric lifetime (year)
CFC-11	CCl ₃ F	4660	1	50
CFC-12	CCl ₂ F ₂	10,200	1	100
Halon-1301	CBrF ₃	6290	10	65
HCFC-22	CHF ₂ Cl	1760	0.055	12.1
HCFC-123	CF ₃ CHCl ₂	79	0.02	1.4
HFC-23	CHF ₃	12,400	0.0	264
HFC-134a	CH ₂ FCF ₃	1300	0.0	14.6
HFC-152a	CH ₃ CHF ₂	138	0.0	1.5
HFC-245fa	CHF ₂ CH ₂ CF ₃	858	0.0	7.6

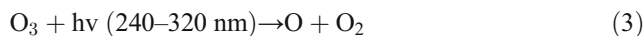
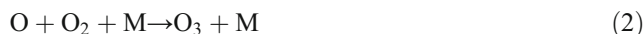
Global warming potential is a measure of the degree of radiative forcing of a given molecule compared to a molecule of CO₂, whose GWP value is standardized to 1 (see footnote 5 for explanation on the 100-year GWP). Ozone depletion potential is the ratio of the impact on O₃ of a chemical compared to the impact of a similar mass of CFC-11, whose ODP value is standardized to 1. Sources: IPCC Special Report on Emission Scenarios, Section 5.4.3. Halocarbon and other halogenated compounds (http://www.grida.no/publications/other/ipcc_sr/?src=/climate/ipcc/emission/); United Nations Environment Programme, OzonAction Branch (http://www.unep.fr/ozonaction/topics/hcfc_list.htm); United States Environmental Protection Agency, Ozone Layer Protection—Science (<http://www.epa.gov/ozone/science/ods/index.html>); IPCC 5th Assessment GWPs Table 8.A.1 (Chapter 8, Appendix 8.A) http://www.climatechange2013.org/images/uploads/WGIAR5_WGI-12Doc2b_FinalDraft_Chapter08.pdf

in the troposphere is harmful to human health and damages vegetation and aquatic ecosystems; ozone in the stratosphere filters out (by absorbing) the UV-B portion of sunlight and protects life on earth from harmful radiation. Most ozone is concentrated in a stratospheric layer between ~15 and 30 km above the Earth's surface. This ozone layer acts as a powerful UV shield, without which complex life forms might not have evolved on Earth. With a series of striking scientific discoveries in the 1970s and 1980s, including the catalytic ozone destruction reactions and the observation of the Antarctic ozone hole, the details of ozone-depleting mechanisms, involving sophisticated atmospheric chemistry and physics, were quickly established and accepted by the scientific community as well as policy makers. Starting 1987, the Montreal Protocol and its several adjustments and amendments were internationally adopted and rapidly implemented.³ The Montreal Protocol is probably the most successful global environmental treaty, based on sound science and moving as fast as technically and economically feasible (several other papers in this Mini Symposium elaborate on the metrics of this success). As a result of this global effort, the stratospheric ozone is expected to return to its 1980 levels by mid-twenty-first century (WMO 2014b).

³ Amendments had new chemicals added to the lists of controlled substances. Adjustments accelerated the schedule for phase-out. Decisions allow continued use of some ODSs for time-limited periods for applications essential or critical, such as medicine or national security.

Stratospheric ozone: depletion by catalytic chain reactions

Paradoxically, atmospheric ozone—acting as the UV shield for the Earth—is created by UV photons. The oxygen molecule O₂ absorbs high-energy, far-UV photons (<242 nm) and is split into two oxygen (O) atoms, each reacting with another O₂ molecule to form O₃. The overall process thus converts three O₂ molecules to two O₃ molecules. The O₃ molecules react with other stratospheric atoms and molecules (often with photons involved), some natural (such as O atoms) and some anthropogenic (such as chlorine (Cl) atoms from CFCs and bromine (Br) atoms from halons). Steady-state conditions lead to the highest concentration of ozone residing in the ozone layer, with the peak concentration at ~25 km in altitude (Seinfeld and Pandis 2006). Above this layer, ozone decreases in concentration mainly due to the decline in O₂ molecules; below, ozone also decreases in concentration because of the decrease in the UV flux needed for O₂ photodissociation. The Chapman Mechanism characterizes these processes of ozone production and destruction as follows (M is a third molecule such as N₂; hv represents photons of different wavelength):



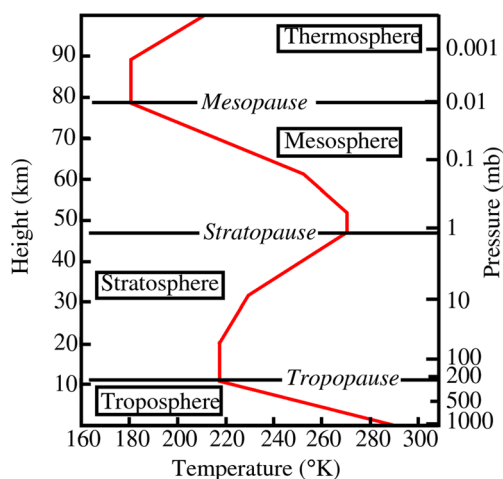
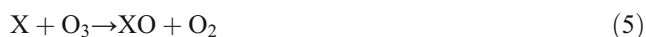


Fig. 1 Layers of the atmosphere (courtesy of Professor Robert Houze in the Department of Atmospheric Sciences at the University of Washington)



However, kinetic calculations and field measurements suggest that the Chapman Mechanism is not sufficient to explain the observed ozone profiles in the stratosphere. Through stunning experimental and theoretical work, a number of atmospheric scientists (e.g., Bates and Nicolet 1950; Crutzen 1970; Johnston 1971; Molina and Rowland 1974; Farman et al. 1985; Solomon et al. 1986) quickly revealed that distinct catalytic chain reactions have to be taken into account in explaining stratospheric ozone depletion, and some of these reactions involve chemicals with anthropogenic origins. The details of these mechanisms and relevant kinetics are thoroughly discussed by Seinfeld and Pandis (2006). In essence, O_3 is converted to O_2 by a chain carrier X (reaction 5), which is restored in the process (reaction 6):



Reactions 5 and 6 provide extra pathways for ozone destruction in addition to reactions 3 and 4 above, yet only a small number of X radicals are needed for this catalytic cycle to repeat many times until X is removed by other processes. Four chemical species can act as the catalyst X: hydroxyl radical (OH), Cl, and Br atoms, as well as nitric oxide (NO). In this paper, the “dot” commonly used to indicate a “free radical” is omitted.

1. $\text{X}=\text{OH}$: OH is formed by the reaction of a water molecule with an excited-state oxygen atom O (originating from reaction 3 with ~2 % yield). CH_4 can also participate in radical reactions to form additional OH radicals,

eventually leading to more ozone destruction. Thus, increasing amounts of water and methane due to human activities, upon entering the stratosphere, can lead to enhanced ozone depletion there.

2. $\text{X}=\text{Cl}$ or Br : Natural sources of Cl and Br (such as methyl chloride and methyl bromide from oceanic emissions)⁴ are unlikely to migrate into the stratosphere in large quantities due to their reactions in the troposphere. However, man-made organic compounds containing halogens (F, Cl, Br) were historically produced in vast quantities, and some are eventually emitted and transported into the stratosphere, as atmospheric measurements confirm. These compounds—for example, CFCs and Br-containing halons—have low reactivity in the troposphere, even with the highly reactive OH radical. Ironically, their chemical stability renders them both excellent materials as non-flammable, low-toxicity refrigerants, blowing agents or fire retardants, and disastrous ozone-depleting materials when they successfully “migrate” into the stratosphere. These can also be seen in their atmospheric lifetimes often in the range of a decade or longer (Table 1). Once in the stratosphere, CFCs and halons are attacked by the powerful UV photons, producing large quantities of Cl or Br atoms after the C-halogen bonds are broken and the ozone destruction catalytic cycles are activated (reactions 5 and 6). While the abundance of Br is much lower than Cl in the stratosphere (by about 150 times), Br is about 50 times more efficient than Cl in catalytically depleting ozone for several reasons: Br is released into the stratosphere more rapidly, it is slower to be removed from the stratosphere, and its reservoir species (BrONO_2) is considerably less stable than that of Cl (see below). Taking into account the relative efficiencies, the measurements of these two halogens are often combined into “Equivalent Effective Stratospheric Chlorine (EESC)”. Figure 2 shows the estimates by WMO of the concentration of EESC in the stratosphere in the mid-latitudes, under full compliance with the Montreal Protocol. The EESC level is expected to return to its benchmark 1980 values around 2040 to 2060. As a result, according to model simulations, ozone is expected to return to its 1980 levels around 2025–2040 for the mid-latitudes and the Arctic and around 2045–2060 for the Antarctic, respectively (WMO 2014b).

3. $\text{X}=\text{NO}$: Nitric oxide (NO) radical in the stratosphere mainly comes from the UV photolysis of N_2O . NO then acts as the catalytic chain carrier X to cycle, through NO_2 , in the ozone destruction processes (reactions 5 and 6). In addition, NO_2 also reacts with OH and ClO radicals, forming HNO_3 and ClONO_2 , respectively. While not

⁴ Methyl bromide (CH_3Br) and nitrous oxide (N_2O) are potent ozone-depleting GHGs that originate from both natural sources and human activities.

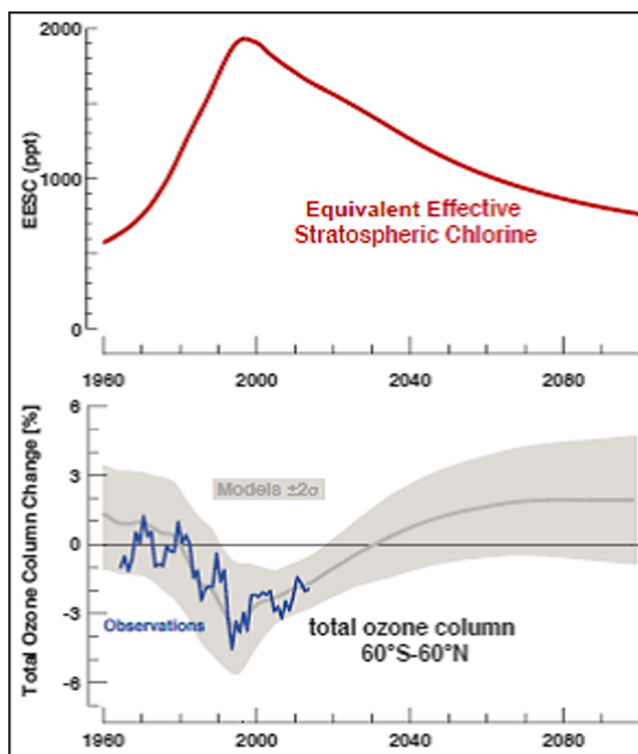
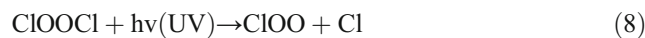
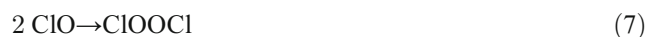


Fig. 2 Top panel variation in EESC at mid-latitudes (1960–2100). The future projection is for the baseline scenario, which assumes complete compliance with the current Montreal Protocol and no further amendments and adjustments. Bottom panel the average total column ozone changes over the same period, from multiple model simulations (gray line, with $\pm 2\sigma$), as well as the observed column ozone changes (1965–2013, blue line). Source: WMO 2014 Global Ozone Research and Monitoring Project, Report No. 56 (http://ozone.unep.org/Assessment_Panels/SAP/SAP2014_Assessment_for_Decision-Makers.pdf)

directly involved in the chain reactions, HNO_3 and ClONO_2 are reservoir species that lock OH and ClO radicals in less reactive forms (thus reducing ozone destruction). Upon interaction with the UV photons, they decompose to release OH and ClO radicals that are again involved in ozone destruction. The NO radical therefore has this unique two-sided effect, either facilitating or inhibiting ozone destruction. Which of the two effects dominates mainly depends on altitude.

The Antarctic ozone hole, referring to the massive ozone loss in the Antarctic springtime, has additional photochemical mechanisms at play that are coupled with special meteorological conditions. Polar stratospheric clouds (PSCs) form in the extremely cold polar vortex in winter. ClONO_2 reacts with HCl and H_2O on cloud droplet surfaces to form the more reactive Cl_2 and HOCl in the dark winter. When sunlight becomes available as spring arrives, photolysis reactions produce a burst of Cl atoms that start the catalytic O_3 destruction process as described earlier (reactions 5 and 6). In addition, ClO radicals form dimers, as shown in reaction 7:



UV photons then dissociate the Cl–O bond to regenerate Cl atoms, as shown in reactions 8 and 9. The rate of reaction 7 is proportional to the square of the ClO concentration, partly explaining the sudden appearance (i.e., fast formation) of the Antarctic ozone hole as Cl atoms are produced rapidly upon the availability of the spring sunlight. While ozone depletion has been observed in both poles, it is less pronounced in the Arctic region since the temperature there does not fall low enough to facilitate ozone destruction.

Tropospheric ozone: warming and pollution

In the troposphere, ozone is formed and destroyed by very different mechanisms. In the remote troposphere, ozone is formed from radical reactions of carbon monoxide or methane involving OH and nitrogen oxides ($\text{NO}_x = \text{NO} + \text{NO}_2$). In the polluted troposphere, ozone is often part of the photochemical smog that causes serious health concerns. The ingredients for smog formation include O_2 , NO_x , volatile organic compounds (VOCs), and sunlight via an interwoven set of photochemical reactions coupled in a nonlinear fashion (Seinfeld and Pandis 2006; Spiro et al. 2012). In essence, ozone is formed by the photolysis of NO_2 , followed by the reaction of O and O_2 . NO_2 , however, needs to be replenished by the reaction of VOCs with OH forming peroxy radicals, which then oxidize NO to form NO_2 . The OH radicals, on the other hand, come from the interaction of UV photons (available in the troposphere despite the stratospheric filtering of UV) with O_3 (reactions 3) followed by the reaction of the O atom with water.

Tropospheric ozone is a significant greenhouse gas, contributing to global warming as much as about 20 % of that due to CO_2 (Molina et al. 2009). Because of the large increase in VOCs, NO_x , CH_4 , and other ozone precursor gases, tropospheric ozone abundance has increased by about 30 % since the preindustrial era (Molina et al. 2009 and references therein). Thus, efforts to reduce tropospheric ozone not only bring public health and food production benefits (reducing harm to humans and crops) but also help mitigate global warming. With a relatively short lifetime (weeks), ground-level ozone reduction has been proposed as a near-term, fast-response climate mitigation strategy (Molina et al. 2009; Wallack and Ramanathan 2009).

CFCs, HCFCs, and HFCs: ozone depleting potential and global warming potential

The paradoxical roles of CFCs and halons—their chemical stability and industrial applications as well as their ability to destroy ozone after migrating into the stratosphere—were discussed in the last section. The Montreal Protocol seeks to phase out CFCs and halons due to their relatively high ODP, as seen in Table 1. HCFCs have generally lower ODP compared to CFCs. However, large-enough quantities of HCFCs could still deplete ozone significantly; thus, they are viewed as transitional CFC substitutes. The Montreal Protocol amendments call for the phase-out of HCFCs, which are to be replaced by compounds with no chlorine. HFCs that replace some CFCs are safe for the ozone layer and typically have much lower GWPs; for example, HFC-134a has an IPCC AR5⁵ $GWP_{100\text{-year}}$ of 1300, compared to the AR5 $GWP_{100\text{-year}}$ of 10,200 for CFC-12 that it replaced globally in motor vehicle air conditioning.

However, while having zero ODP, some HFCs that replace HCFCs often have equivalent or higher GWP (Table 1), landing them on the list of controlled GHGs for emission reductions under the Kyoto Protocol. With the expected increase in the use of HFCs due to increasing population and demand, global emissions of HFCs in 2050 are projected to be equivalent to 9–19 % ($CO_2\text{-eq}$) of projected CO_2 emissions in business-as-usual scenarios or even higher in a 450-ppm CO_2 -stabilization scenario (Velders et al. 2009). Thus, the contribution to global warming by HFCs should not be ignored in climate policy making. The phase-down of HFCs has been proposed as a near-term mitigation measure, and its inclusion in the Montreal Protocol is now under international negotiation (Velders et al. 2009; Molina et al. 2009; Shoemaker et al. 2013).

Other short-lived climate forcers

In addition to HFCs and tropospheric O_3 , two other SLCFs are also being considered as near-term mitigation options: CH_4 and BC. Emission cuts of either of these two species can potentially slow down global warming with a fast response, as discussed in this special issue (Canan et al. 2014; Zaelke and Borgford-Parnell 2014). Similar to CO_2 and N_2O , global CH_4 concentration has increased dramatically since the pre-industrial era due

to human activities, contributing significantly to global warming via greenhouse effects (IPCC 2013). As mentioned earlier, CH_4 also contributes to tropospheric O_3 formation, further contributing to global warming. Thus, reductions in CH_4 emissions have the potential for both climate mitigation (less CH_4 -induced and O_3 -induced warming) and pollution control (less O_3 -induced pollution). However, the enthusiasm for methane control appears to be less than the other three SLCFs at this point. The lifetime of methane, about a decade, is much longer than BC and tropospheric ozone. In addition, the required actions in the agriculture, energy, and waste treatment sectors pose huge operational challenges (Ramanathan and Xu 2010).

By comparison, reducing BC has been discussed with much fanfare (Shoemaker et al. 2013; Burney et al. 2013; Shindell et al. 2012; Molina et al. 2009), including in the foreign policy arena (Wallack and Ramanathan 2009), despite the lack of a full understanding of its net climate effects (Bond et al. 2013; Chen et al. 2010). Black carbon is a major component of soot and is produced by incomplete combustion of fossil fuels and biomass. Its emission sources include diesel vehicles, residential stoves, forest fires, agricultural open burning, and some industrial facilities. BC affects global radiative forcing both directly by absorbing solar radiation and indirectly via the modification of cloud lifetime and albedo (Chung and Seinfeld 2005; Jacobson 2001). BC may also change the radiation balance if it falls on ice or snow, causing faster melting and having further feedback effects. In the scenarios where BC is mixed with sulfate and/or organic aerosols, the net aerosol forcing needs to be carefully evaluated. Overall, substantial uncertainties exist as to the net radiative forcing of BC-containing aerosols, in particular due to the lack of knowledge regarding the interactions of clouds with both BC and co-emitted organic carbon (Bond et al. 2013).

Owing to uncertainties in the science as well as the implementation, climate benefits from reductions in SLCFs could be smaller than the highest estimates and may maximize when reductions in LCLFs are concurrently implemented (Smith and Mizrahi 2013; Bowerman et al. 2013). Nevertheless, managing these SLCFs is an attractive option in partially, but quickly, curbing global warming, thus delaying and helping avoid abrupt climate change (Molina et al. 2009; Shoemaker et al. 2013). In addition, modeling studies suggest that mitigating the SLCFs can have a significant impact on sea level rise (SLR), reducing the cumulative SLR by 22–42 % by 2100 (Hu et al. 2013). While the science continues to be investigated, there can be manageable and affordable measures to take with immediate benefits, such as replacing the kerosene lamps that are widely used as light sources in developing countries. The emissions from these sources are nearly all black carbon, exerting a positive (warming) climate forcing (Lam et al. 2012).

⁵ The latest estimates of GWP are from the 2013 Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5). GWP is typically expressed for a 100-year time horizon ($GWP_{100\text{-year}}$), but GHGs with atmospheric lifetime longer than 100 years continue to exert a climate forcing for up to 10,000 years (for example, Perfluorocarbons—PFCs).

Interplay of climate change and ozone depletion: making science-based policies

While climate change (most pointedly reflected in global warming) and ozone depletion are largely two separate phenomena with different mechanisms, they share common components that render them not entirely independent of each other.

In some cases, one problem exacerbates the other. For example, with global warming, there is more water vapor in the atmosphere. CH₄ emissions also continue to rise due to human activity. In addition to causing further warming (feedback effects), both water and methane molecules migrate into the stratosphere, where they can contribute to more production of the hydroxyl radical, which in turn causes further ozone depletion. Another example is the possibility of a more severe ozone hole in the Arctic, arising from the further cooling in the stratosphere in keeping with the ongoing tropospheric warming, as has been observed (both the increase in GHGs and the stratospheric ozone loss play key roles in the stratospheric cooling, which is not yet fully understood).

However, recognizing this interplay and implementing science-based policies to solve environmental problems in tandem (co-benefits approach) would be more desirable for preserving our planetary sustainability. The Montreal Protocol has set an extraordinary example by mitigating stratospheric ozone depletion while also bringing substantial climate benefits. It is estimated that, thanks to the Protocol, the radiative forcing from ODSs reached in 2000, and has since remained, at around 0.32 W/m². Without the Protocol, however, this forcing could have reached 0.60 to 0.65 W/m², or about 40 % of that of CO₂ in 2010, and the world would have been warmer (Velders et al. 2007, 2009 and 2012). Knowing that the concentrations of HFCs are on the track of fast increase in the coming decades, which exacerbates global warming, limiting their use under a further-amended Montreal Protocol is worthy of consideration by the international community. More broadly, measures of limiting SLCFs, including also tropospheric ozone, black carbon, and methane, are worthy of consideration on a global scale. Unequivocally, each of the last three decades has been successively warmer at the Earth's surface than any preceding decade since 1850 (IPCC 2013). Even a modest slowdown in global warming shall win time and hopefully political will to build international coalitions and implement novel technologies in curbing climate change by eventually limiting LLCFs such as carbon dioxide.

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