

Black Carbon and Its Effects on Climate

2.1 Summary of Key Messages

- Black carbon (BC) is the most strongly light-absorbing component of particulate matter (PM), and is formed by the incomplete combustion of fossil fuels, biofuels, and biomass.
 - BC can be defined specifically as a solid form of mostly pure carbon that absorbs solar radiation (light) at all wavelengths. BC is the most effective form of PM, by mass, at absorbing solar energy. BC is a major component of “soot”, a complex light-absorbing mixture that also contains organic carbon (OC).
 - Other carbon-based PM may also be light-absorbing, particularly brown carbon (BrC), which is a class of OC compounds that absorb light within the visible and ultraviolet range of solar radiation and that can exist within the same particles as BC. The net contribution of BrC to climate is presently unknown.
- BC is always emitted with other particles and gases, such as sulfur dioxide (SO₂), nitrogen oxides (NO_x), and OC. Some of these co-emitted pollutants exert a cooling effect on climate. Therefore, estimates of the net effect of BC emissions sources on climate should include the offsetting effects of these co-emitted pollutants.
- Atmospheric processes that occur after BC is emitted, such as mixing, aging, and coating, can also affect the net influence of BC on climate.
- The short atmospheric lifetime of BC (days to weeks) and the mechanisms by which it affects climate distinguish it from long-lived greenhouse gases (GHGs) like carbon dioxide (CO₂).
 - Targeted strategies to reduce BC emissions can be expected to provide climate responses within the next several decades. In contrast, reductions in GHG emissions will take longer to influence atmospheric concentrations and will have less impact on climate on a short timescale, but deep reductions in GHG emissions are necessary for limiting climate change over the long-term.
- Emissions sources and ambient concentrations of BC vary geographically and temporally, resulting in climate effects that are more regionally and seasonally dependent than the effects of long-lived, well-mixed GHGs. Likewise, mitigation actions for BC will produce different climate results depending on the region, season, and emissions category.
- BC influences climate through multiple mechanisms:
 - *Direct effect:* BC absorbs both incoming and outgoing radiation of all wavelengths, which contributes to warming of the atmosphere and dimming at the surface. In contrast, GHGs mainly trap outgoing infrared radiation from the Earth’s surface.
 - *Snow/ice albedo effect:* BC deposited on snow and ice darkens the surface and decreases reflectivity (albedo), thereby increasing absorption and accelerating melting. GHGs do not directly affect the Earth’s albedo.
 - *Other effects:* BC also alters the properties and distribution of clouds, affecting cloud reflectivity and lifetime (“indirect effects”), stability (“semi-direct effect”), and precipitation. These impacts are associated with all ambient particles, but not GHGs.
- The direct and snow/ice albedo effects of BC are widely understood to lead to climate warming. Based on the studies surveyed for this report, the direct and snow/ice albedo effects of BC together likely contribute more to current warming than any GHG other than CO₂ and methane (CH₄).
- The climate effects of BC via interaction with clouds are more uncertain, and their net climate influence is not yet clear.

- All aerosols (including BC) affect climate indirectly by changing the reflectivity and lifetime of clouds. The net indirect effect of all aerosols combined is very uncertain but is thought to have a net cooling influence. The contribution of BC to this cooling has not been quantified.
- BC has additional effects on clouds—including changes to cloud stability and enhanced precipitation from colder clouds—that can lead to either warming or cooling.
- The net climate influence of these cloud interaction effects of BC is not yet clear. There is inconsistency among reported observational and modeling results, and many studies do not provide quantitative estimates of cloud impacts.
- The sign and magnitude of the net climate forcing from BC emissions are not fully known at present, largely due to remaining uncertainties regarding the effects of BC on clouds. Though most estimates indicate that BC has a net warming effect, a net cooling influence cannot be ruled out. Further research and quantitative assessment are needed to reduce remaining uncertainties.
- Regional climate impacts of BC are highly variable, and sensitive regions such as the Arctic and the Himalayas are particularly vulnerable to the warming and melting effects of BC. Estimates of snow and ice albedo forcing in key regions also exceed global averages.
- BC also contributes to the formation of Atmospheric Brown Clouds (ABCs) and resultant changes in the pattern and intensity of precipitation.
- Due in large part to the difference in lifetime between BC and CO₂, the relative weight given to BC as compared to CO₂ (or other climate forcers) in terms of its impact on climate is very sensitive to the formulation of the metric used to make the comparison.
- There is currently no single metric that is widely accepted by the science and research community for this purpose.
- There are several metrics that have been applied to the well-mixed GHGs with respect to different types of impacts, especially the global warming potential (GWP) and global temperature potential (GTP). These metrics can be applied to BC, but with difficulty due to important differences between BC and GHGs. Recently, new metrics designed specifically for short-lived climate forcers like BC have been developed, including the specific forcing pulse (SFP) and the surface temperature response per unit continuous emission (STRE).
- There is significant controversy regarding the use of metrics for direct comparisons between the long-lived GHGs and the short-lived particles for policy purposes; however, these comparisons are less controversial when used for illustrative purposes.
 - There are a number of factors that should be considered when deciding which metric to use, or whether comparisons between BC and CO₂ are useful given a particular policy question. These include: the time scale (e.g., 20 years, 100 years, or more), the nature of the impact (radiative forcing, temperature, or more holistic damages), the inclusion of different processes (indirect effects, snow albedo changes, co-emissions), and whether sources and impacts should be calculated regionally or globally.
 - If the primary goal is reducing long-term change, then a metric like a 100-year GWP or GTP would be more appropriate. If the rate of near-term climate change and near-term damages to sensitive regions like the Arctic are also a consideration, there is no single existing metric that adequately weights impacts over both time periods, and a multi-metric approach may be more appropriate than developing a single metric that attempts to serve all purposes.

2.2 Introduction

There is a general consensus within the scientific community that BC is contributing to climate change at both the global and regional levels. Like CO₂, BC is produced through the burning of carbon-based fuels, including fossil fuels, biofuels and biomass. BC is part of the mix of PM released during the incomplete combustion of these fuels. BC influences climate by absorbing sunlight when suspended in the atmosphere or when deposited on the Earth's surface. The energy absorbed by BC is then released as heat and contributes to atmospheric warming and the accelerated melting of ice and snow. In addition, BC is capable of altering other atmospheric processes, such as cloud formation and evaporation, and precipitation patterns.

The strong absorption, short atmospheric lifetime, and other characteristics of BC make its impacts on climate different from those of long-lived GHGs like CO₂ (see Figure 2-1). Because BC is involved in complex atmospheric physical and chemical processes, it is difficult to disentangle all associated impacts and to evaluate its net effect on climate. In addition, the combustion processes that produce BC also produce other pollutants, such as SO₂, NO_x, and OC. Since many of these compounds have a cooling effect, BC's impacts are mixed with—and sometimes offset by—these co-emitted substances. This must

be considered when evaluating the net effect of emissions sources.

This chapter focuses on how and to what extent BC influences the Earth's climate. Specifically, this chapter discusses approaches for defining BC and other light-absorbing particles, highlights the differences between BC and GHGs, and addresses the role of co-emitted pollutants. Further, this chapter summarizes recent scientific findings regarding the processes by which BC affects climate and the magnitude of BC's impacts on global and

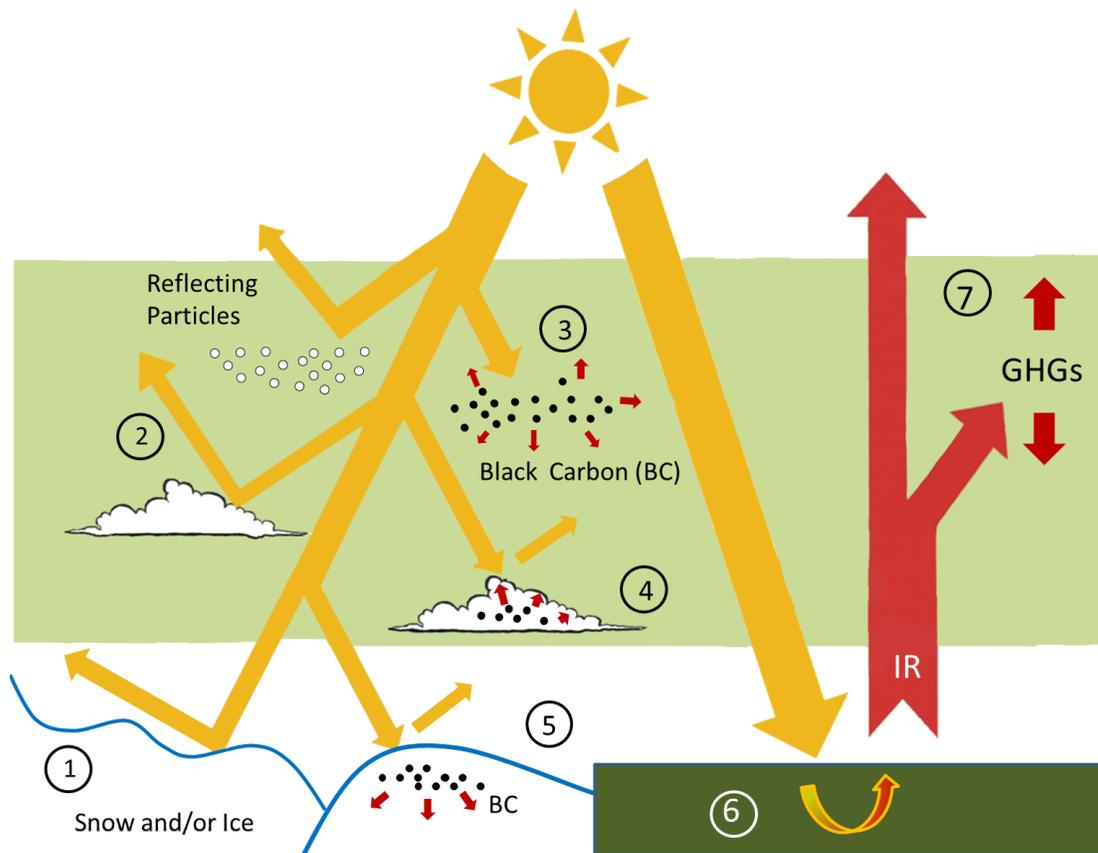


Figure 2-1. Effects of BC on Climate, as Compared to GHGs. (Source: U.S. EPA)

1. Sunlight that penetrates to the Earth's surface reflects off bright surfaces, especially snow and ice.
2. Clean clouds and non-light-absorbing (transparent) particles scatter or reflect sunlight, reducing the amount of solar energy that is absorbed by the surface.
3. BC suspended in the atmosphere absorbs some incoming solar radiation, heating the atmosphere.
4. Clouds containing BC inclusions in drops and BC interstitially between drops can absorb some incoming solar radiation, reducing the quantity that is reflected. Clouds warmed by the absorbed energy have shorter atmospheric lifetimes and may be less likely to precipitate compared to clean clouds.
5. BC deposited on snow and/or ice absorbs some of the sunlight that would ordinarily be reflected by clean snow/ice, and increases the rate of melting.
6. Most solar radiation is absorbed by the Earth's surface and warms it. Part of the absorbed energy is converted into infrared radiation that is emitted into the atmosphere and back into space.
7. Most of this infrared radiation passes through the atmosphere, but some is absorbed by GHG molecules like CO₂, methane, ozone and others. These gases re-emit the absorbed radiation, with half returning to the Earth's surface. This GHG effect warms the Earth's surface and the lower atmosphere.

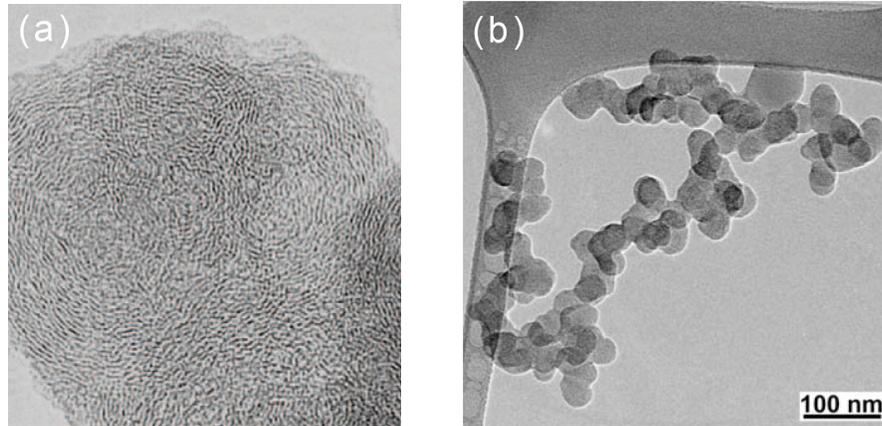


Figure 2-2. BC Images. (a) High resolution transmission electron microscopy (TEM) image of a BC spherule (Pósfai and Buseck, 2010). (b) TEM image of a representative soot particle. Freshly emitted soot particles are aggregates of soot spherules (Alexander et al., 2008).

regional climate, highlighting the effect of BC in sensitive regions such as the Arctic and other snow- and ice-covered regions. The chapter discusses the significant remaining uncertainties about BC's effects on climate, and the need for further research in key areas. The final section of this chapter introduces several metrics that can be used to quantify the climate impacts of BC and other pollutants (such as CO₂ and CH₄) relative to a common baseline. The section highlights the fact that there is no one "best" metric for comparing BC to other pollutants and that the utility of each metric depends on the policy objective.

2.3 Defining Black Carbon and Other Light-Absorbing PM

All PM in the atmosphere can affect the Earth's climate by absorbing and scattering light. Sunlight absorbed by PM increases the energy in the Earth's climate system, leading to climate warming. Conversely, light scattered by PM generally leads to increased reflection of light back to space, leading to climate cooling (Charlson, 1992; Moosmüller et al., 2009; Seinfeld and Pandis, 2006; Forster et al., 2007). Carbonaceous PM, a class of material found in primary and secondary particles, has typically been divided into two classes: BC and OC (see text box on "Terminology"). Neither BC nor OC has a precise chemical definition. The term BC generally includes the solid forms of carbon emitted by incomplete combustion while OC refers to the complex mixtures of different carbon compounds found in both primary and secondary carbonaceous particles. Carbonaceous PM includes an array of organic compounds that, along with BC, possess radiative properties that fall along a continuum from light-

absorbing to light-scattering. Both BC and OC are part of the broader category of suspended particles and gases known as aerosols, all of which have light-absorption and light-scattering properties.

In this report, BC is defined as the carbonaceous component of PM that absorbs all wavelengths of solar radiation.¹ For this reason, among the many possible forms of PM, BC absorbs the most solar energy. Per unit of mass in the atmosphere, BC can absorb a million times more energy than CO₂ (Bond and Sun, 2005), making it a significant climate warming pollutant in regions affected by combustion emissions.

BC forms during combustion, and is emitted when there is insufficient oxygen and heat available for the combustion process to burn the fuel completely (see text box on "Products of Incomplete Combustion"). BC originates as tiny spherules, ranging in size from 0.001 to 0.005 micrometers (μm), which aggregate to form particles of larger sizes (0.1 to 1 μm) (Figure 2-2). Particles in this range are similar in size to the wavelengths emitted by the sun, making them especially effective in scattering or absorbing these wavelengths (Horvath, 1993). The characteristic particle size range in which fresh BC is emitted also makes it an important constituent of the ultrafine (<100 nanometers (nm)) subclass of PM_{2.5}.

¹The spectrum of solar radiation striking Earth's atmosphere ranges from high energy UV with wavelengths shorter than 280 nm down to infrared radiation as long as 1000 nm. However, UV wavelengths shorter than 280 nm are substantially absorbed by the stratosphere. For the purposes of this discussion, the term "all wavelengths of solar radiation" corresponds to the solar wavelengths present in the troposphere (e.g., in the range 280 - 2500 nm).

Terminology

Black carbon (BC) is a solid form of mostly pure carbon that absorbs solar radiation (light) at all wavelengths. BC is the most effective form of PM, by mass, at absorbing solar energy, and is produced by incomplete combustion.

Organic carbon (OC) generally refers to the mix of compounds containing carbon bound with other elements like hydrogen or oxygen. OC may be a product of incomplete combustion, or formed through the oxidation of VOCs in the atmosphere.² Both primary and secondary OC possess radiative properties that fall along a continuum from light-absorbing to light-scattering.

Brown carbon (BrC) refers to a class of OC compounds that absorb ultraviolet (UV) and visible solar radiation. Like BC, BrC is a product of incomplete combustion.³

Carbonaceous PM includes BC and OC. Primary combustion particles are largely composed of these materials.

Light absorbing carbon (LAC) consists of BC plus BrC.

Soot, a complex mixture of mostly BC and OC, is the primary light-absorbing pollutant emitted by the incomplete combustion of fossil fuels, biofuels, and biomass.

BC is emitted directly from sources, making it a form of *primary* PM. This distinguishes it from *secondary* PM such as sulfates, nitrates and some forms of OC that are formed in the atmosphere from gaseous precursors like SO₂, NO_x and volatile organic compounds (VOCs).

When BC is emitted directly from sources as a result of the incomplete combustion of fossil fuels, biofuels and biomass, it is part of a complex particle mixture called *soot* which primarily consists of BC and OC. This mixture is the light-absorbing component of these air pollution emissions.

Soot mixtures can vary in composition, having different ratios of OC to BC,² and usually include inorganic materials such as metals and sulfates. For example, the average OC:BC ratio among global sources of diesel exhaust is approximately 1:1. For biofuel burning, the ratio is approximately 4:1 and for biomass burning it is approximately 9:1

²When referring to emissions and measurements, OC denotes the total carbon associated with the organic compounds, while organic mass (OM) refers to the mass of the entire carbonaceous material, including hydrogen and oxygen. Similarly, measurements and emissions reported as elemental carbon (EC) denote the non-organic, refractory portion of the total carbon and is an indicator for BC. For more details, see Chapter 5 and Appendix 1.

(Lamarque et al., 2010). As expected, very dark soot indicates the presence of low OC:BC ratios. As the OC fraction begins to dominate, the color of the soot mixture shifts to brown and yellow. A brown soot sample is dominated by a form of OC known, as might be expected, as "brown carbon" (BrC).³ BrC, another product of incomplete combustion, absorbs portions of the visible spectrum, but is less effective in capturing solar energy than BC (Alexander et al., 2008; Novakov and Corrigan, 1995b). The mixture shifts in color toward yellow when the emissions source is no longer producing BC and BrC. Yellow carbon, another form of OC, is also able to absorb visible radiation, but to a lesser extent than BrC (Bond, 2001; Gelencsér, 2004; Andreae and Gelencsér, 2006). Figure 2-3 illustrates the variance in soot composition resulting from different fuels and stages of fuel combustion. The stages of fuel combustion responsible for producing BC and the various forms of OC observed in soot are described in the text box on this page.

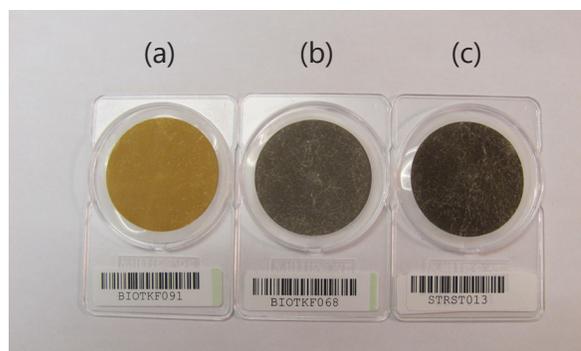


Figure 2-3. Representative Examples of Filter Samples Collected from Different Sources, including: (a) Smoldering Biomass, (b) Flaming Biomass, and (c) Diesel Exhaust. (Photo courtesy of Desert Research Institute)

In general, light absorption by carbonaceous PM can be described as a continuum from light-absorbing to light-scattering with BC at one end, most OC at the other, and BrC occupying the partially absorbing

³During solid fuel combustion, BrC forms during the preheating (pyrolysis) phase, and during both flaming and smoldering combustion. The light-colored smoke characteristic of the pyrolysis and smoldering combustion phases is primarily OC, including both BrC and other forms of OC, and does not include soot. Secondary BrC can also form during reactions, similar to polymerization, that take place in primary particles as emissions plumes age. BrC of this type is known as "humic-like substances" (HULIS).

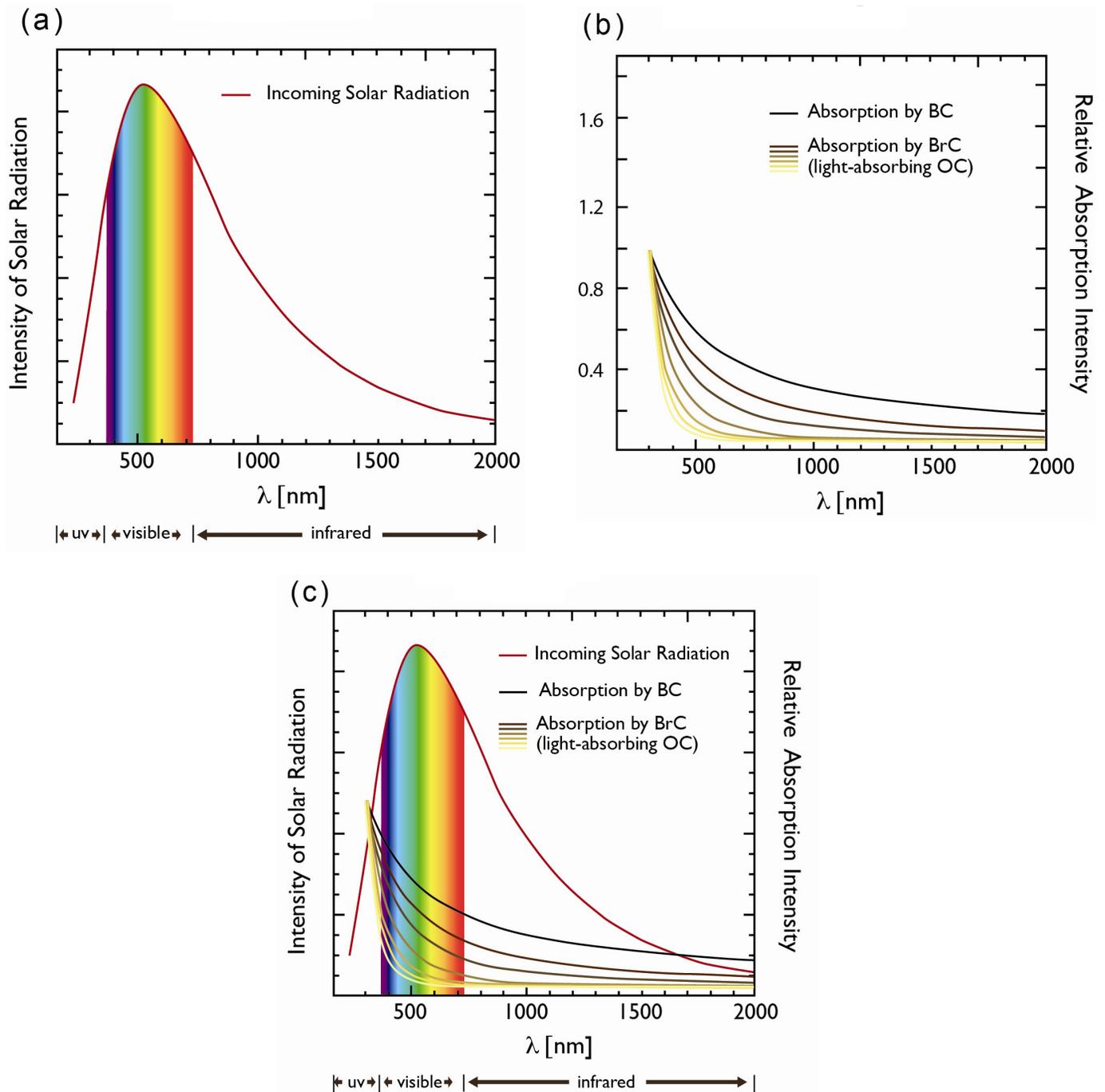


Figure 2-4. Light Absorption by BC, BrC, and Ambient Mixtures. (Source: U.S. EPA)

a. The radiation wavelengths emitted by the Sun that reach the Earth begin around 280 nm (UV-B), peak in the mid-visible range, and reach out past 2000 nm (Infrared). The shorter the wavelength, the higher its energy.

b. The extent to which BC and BrC absorb solar radiation depends upon the wavelength of incoming light. This plot shows idealized examples of those dependencies, assuming that both forms of carbon absorb to the same extent at 280 nm. BC is more effective in absorbing solar energy across the entire solar spectrum than any form of BrC. BrC increasingly declines in its capacity to absorbing longer wavelengths as the mixture moves from larger light-absorbing compounds to smaller compounds, as indicated by the change in color from dark brown to yellow.

c. When the light absorption curves for BC and BrC are superimposed upon the solar spectrum, the significance of the different absorption efficiencies between BC and BrC becomes evident. BC will, all else being equal, absorb more total solar radiation than BrC. In practice, the mass ratios of BC and BrC, along with the specific composition of the BrC mixture, determine the degree to which each form of carbon absorbs the solar energy penetrating the emissions plume.

Products of Incomplete Combustion

Most combustion occurring on Earth (both anthropogenic and natural) involves carbon-based fuels, including fossil fuels (e.g., coal, oil, and natural gas), biomass (e.g., wood and crop residues), and biofuels (e.g., ethanol). Complete combustion of a carbon-based fuel means all carbon has been converted to CO₂. Once ignited, the fuel must be well mixed with oxygen at a sustained high temperature for this to occur. Incomplete combustion emits various materials in both gas and particle form, depending on the combustion conditions (e.g., oxygen availability, flame temperatures, and fuel moisture) and the type of fuel burned (e.g., gas, liquid, or solid). The PM emissions are generically known as soot.

BC is formed during the **flaming** phase of the combustion process. The quantity of BC emitted depends largely on combustion conditions. If there is sufficient oxygen and high temperatures, the soot will be completely oxidized, and BC emissions will be minimal. To increase fuel efficiency and reduce soot emissions, closed combustion systems (e.g., furnaces, combustors, reactors, boilers, and engines) are engineered to increase the mixing of air with the fuel and are insulated to ensure temperatures remain high. Open and uncontrolled burning produces large quantities of BC because oxygen availability and temperatures within the fire can vary widely.

The form of the fuel also influences the likelihood of complete combustion:

- *Gas phase fuels* (e.g., natural gas) can be readily mixed with oxygen, which reduces the emission of carbonaceous particles.
- *Liquid fuels* (e.g., gasoline) generally must vaporize in order to fuel flaming combustion. If a liquid fuel contains heavy oils, vaporization and thorough mixing with oxygen are difficult to achieve. The heavy black smoke emitted by some marine vessels (which burn a sludge-like grade of oil known as “bunker fuel”) is evidence of substantial BC emissions.
- *Solid fuels* (e.g., wood) require preheating and then ignition before flaming combustion can occur. High fuel moisture can suppress full flaming combustion, contributing to the formation of BrC particles as well as BC (Graber and Rudich, 2006; Pósfai et al., 2004; Alexander et al., 2008).

Thermal breakdown of high molecular weight fuels, known as **pyrolysis**, produces a wide array of BrC compounds. When sustained, pyrolysis converts solid fuels such as coal and biomass into char, while releasing volatile gases that can fuel flaming combustion. There is also a non-flaming process known as **smoldering** that is a slower, cooler form of combustion which occurs as oxygen directly attacks the surface of heated solid fuel. The smoke that appears is light-colored, consisting of a variety of organic particles composed of BrC. BC does not form under these conditions, since temperatures are too low to sustain flaming combustion. During open or uncontrolled burning of solid fuels, all stages of the burning process—pyrolysis, smoldering, and flaming combustion—occur simultaneously, in different parts of the fuel pile, resulting in emissions of both BC and BrC.

middle ground.⁴ The characteristic light absorption spectra vary significantly among individual BrC compounds, but are almost entirely limited to the UV to visible portion of the solar spectrum (Jacobson, 1999). Mixtures of these compounds range in color from yellow to brown, roughly corresponding to the average molecular weight of the light-absorbing compounds present (see Figure 2-4). Emissions dominated by smaller BrC compounds will appear yellow, while plumes containing high concentrations

of heavy polycyclic aromatic hydrocarbons (PAHs) will appear darker brown.

BC and different mixtures of BrC show different patterns of light absorption versus wavelength. Light absorption by BC tends to decline more slowly with increasing wavelength, while the falloff in absorption by BrC is always faster than that of BC, tending to vary depending on the composition of the BrC mixture. Absorption by the aforementioned yellow BrC mixtures, dominated by lower molecular weight compounds, falls off very quickly. Dark brown BrC mixtures containing large PAHs or HULIS (see footnote 3) continue absorbing to a significant degree across the solar spectrum.

BrC typically accompanies BC in soot particles. However, independent BrC particles may form in the uncontrolled burning conditions typical of biomass burning, or during inefficient combustion of biofuels. These independent particles, labeled “tar

⁴ Particles containing iron and other calcium, aluminum, and potassium oxides also absorb light. Like BrC, metal oxides are very effective, more so than BC, at absorbing light at shorter wavelengths. Some metal oxides are derived from heavy fuel sources such as residual fuel oil (Huffman et al., 2000). High concentrations of such particles can result from windblown dust and may be significant during dust and sand storms that occur in Africa, China, and the Middle East. These fine particle constituents can travel long distances and may contribute to a positive radiative forcing to a limited degree (Prospero et al., 2010; Liu et al., 2008a).

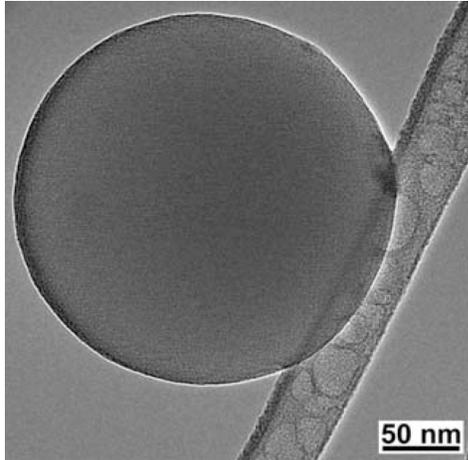


Figure 2-5. TEM Image of a BrC Particle. These particles are referred to as “tar balls” or “carbon spheres” in the literature. (Alexander et al., 2008)



Figure 2-6. Coarse Urban PM (Diameter > 2.5 microns) with a Black Surface Coating, shown on a Filter from a Harvard Impactor Run Approximately 100 feet Above Street Level in Boston. (Photo courtesy of NESCAUM)

balls” or “carbon spheres,” range in diameter from 0.03 and 0.5 μm (Pósfai et al., 2004; Chakrabarty et al., 2010; Adachi and Buseck, 2008). Alexander et al. (2008), in their study of East Asian pollution plumes, found numerous examples of these particles (see Figure 2-5). They noted that, on the basis of their statistical analysis, these particles are very abundant.

Fuel type and burning conditions determine the quantity of BrC produced from a particular combustion source. Some sources, such as open biomass burning, can produce substantially more BrC than BC. On the basis of quantity alone, BrC may lead to greater total solar energy absorbed than BC for those sources, despite the fact that BrC absorbs less energy than BC per unit mass.

Until recently, most measurements of light absorbing carbonaceous PM focused on BC, classifying all other carbon as OC. Growing awareness of the presence of light-absorbing BrC in biomass emissions has prompted the recent suggestion from the scientific community that absorption by BrC should be explicitly included in the evaluation of the role of LAC in climate warming, by accounting for emissions of BrC, along its known radiative properties, in climate modeling studies. Unfortunately, those data do not yet exist.

Combustion emissions, depending upon the purity of the fuel and burning conditions, can also contain a number of inorganic pollutants. For example, combustion of high-sulfur coal is a well-known source of SO_2 and sulfuric acid emissions. Mineral

salts, such as potassium chloride, are emitted during biomass burning due to the presence of potassium in wood and other plant materials. Localized regions of very high heat within a combustion mixture produce NO_x , which further reacts to form nitrates. These co-pollutants in a combustion plume can subsequently form light-scattering particles. The light scattered by these particles may offset the warming effect due to the light absorbed by BC and BrC in the emissions plume. This effect is discussed further later in this chapter in sections 2.5 and 2.6.1.5.

The focus of the discussion up to this point has been on the light absorbing properties of carbonaceous PM (i.e., of the classes of material that appear in particle form in the atmosphere). The chemical composition and physical structure of *individual particles* are also factors determining the overall radiative properties of an emissions plume. Size and morphology influence the efficiency of light absorption by a particle. Particle size and chemical composition change as a fresh emissions plume begins mixing with the ambient atmosphere. Light absorption by BC can be enhanced by 30% to 100% when chemical processing in the atmosphere creates a transparent coating⁵ on the surface of the particle (Fuller et al., 1999; Shiraiwa et al., 2009; Bond et al., 2006a). These effects and the role they play in the overall radiative effect of a combustion plume on regional climate are discussed in section 2.5.

⁵Coatings effective in enhancing light absorption by BC include sulfuric acid from the oxidation of SO_2 , water from cloud-processing and secondary OC that does not absorb solar radiation.

There is also a significant BC component in (or rather on) coarse particles (PM_{10-2.5} and larger), especially in urban areas where coarse-mode particles (such as from re-entrained road dust) are often coated with BC, as shown in Figure 2-6. This coarse urban PM is black, not earth-colored, and likely results from a BC surface coating of coarse mode particles, rather than from a uniform BC composition. While coarse particles will have a very limited effect on climate, they represent a means of human exposure to BC. Chapter 3 discusses the currently understood effects of coarse particles and BC on human health.

BC has been studied as a component of soils and sediments, in addition to its role in air pollution and climate. It plays an important role in various biological, geochemical processes, and has been used as a marker for local vegetation fire histories (Schmidt and Noack, 2000). BC has also been the subject of intense study in the combustion science and engineering fields (Frenklach, 2002). Given the number of distinct scientific disciplines that have studied BC, and the different chemical and environmental contexts in which it appears, many different measurement and estimation approaches exist in the literature, each with their own operational definitions of BC. In the atmospheric sciences alone, the terms “graphitic carbon”, “apparent elemental carbon” (ECa), “equivalent black carbon” (BCe), “light absorbing carbon”, “carbon black”, “soot” and “black smoke” are used interchangeably with or as surrogates for BC materials (Bond and Bergstrom, 2006; Andreae and Gelencsér, 2006).

Although these commonly used terms are not strictly equivalent, we believe that the validity of our analyses and conclusions are not materially compromised by our adopting the convention of using surrogate measurements for BC and soot. EPA has traditionally used surrogate or indicator measurements for many pollutants, including PM_{2.5} whose current regulatory characterization by EPA is based on stable, historically available, consistent and reproducible measurements (Watson et al., 1995). For this report, we believe that BC or elemental carbon (EC) measurements are the best available indicators of BC and soot as these particles are only directly emitted from incomplete combustion, whereas OC can be derived from several sources (e.g., pollens, spores, condensed vapors, secondary aerosols). Nevertheless, the most commonly used measurements may not fully capture the light absorption by BrC, and thus current emissions estimates (and observations on which they are based) may underestimate the positive radiative forcing associated with these particles. The connections between highly correlated BC and

EC measurements and their physical properties are discussed further in Chapter 5.

For purposes of regional air quality management (e.g., human health studies related to air quality, the evaluation of modeled estimates, and the attribution of emissions to sources), BC is measured as a constituent of ambient PM_{2.5}, and is expressed in units of mass. Moreover, BC emissions inventories, as discussed in Chapter 4, are also generally expressed in mass units (e.g., tons/year).

Thus, light-absorption measurements are often converted into estimates of carbon mass. This practice, however, may also contribute to some of the uncertainty in reported ambient concentrations and emissions estimates. Due to reliance on these mass-based indicators, BC is frequently labeled EC due to the long-standing use of carbon measurement methods from which the air quality and emissions estimates were derived. This issue is also evident in health studies that sometimes make a distinction between BC and the measurement on which it is based.

These issues are discussed more fully in Chapter 5 and Appendix 1, which provide a brief description of the various BC and LAC absorption- and mass-related measurement approaches.

2.4 Key Attributes of BC and Comparisons to GHGs

The net impact of BC on climate depends on a number of other factors in addition to its powerful light-absorption capacity. These include atmospheric lifetime, the geographic location of emissions, altitude, interactions with clouds, the presence of co-emitted pollutants, and the influence of aging and mixing processes in the atmosphere. In many of these aspects, BC differs substantially from long-lived GHGs, as summarized in Table 2-1. These differences have implications for how BC influences climate and the climate benefits of BC mitigation as compared to CO₂ mitigation. Each of these dimensions is explored further below.

Particles in general have relatively **short atmospheric lifetimes** in comparison to GHGs. Particles of any type, including BC, are removed from the atmosphere within days to weeks by precipitation and/or dry deposition to surfaces. This short atmospheric lifetime curtails their total contribution to the Earth’s energy balance, even for those particles like BC that have strong absorptive capacity. The efficiency with which particles are removed is influenced by their size and chemical

Table 2-1. Comparison of BC to CO₂ on the Basis of Key Properties that Influence the Climate. (Source: U.S. EPA)

Property	BC	CO ₂
Atmospheric lifetime	Days to weeks	Up to millennia ^e
Distribution of atmospheric concentrations	Highly variable both geographically and temporally, correlating with emission sources	Generally uniform across globe
Direct radiative properties	Absorbs all wavelengths of solar radiation	Absorbs only thermal infrared radiation
Global mean radiative forcing	+0.34 to 1.0 W m ⁻² direct forcing ^a +0.05 W m ⁻² (snow/ice albedo forcing) ^b ± ? (cloud interactions) ^c Net effect: uncertain, but likely warming	+1.66 (±0.17) W m ^{-2(f)}
Cloud interactions	Multiple cloud interactions that can lead to warming or cooling (typically cooling), as well as effects on precipitation	Increases cloud droplet acidity
Surface albedo effects	Contributes to accelerated melting of snow/ice and reduces reflectivity by darkening snow and ice, enhancing climate warming	No direct surface albedo effects
Contribution to current global warming	Likely third largest contributor (after CO ₂ and CH ₄), but large uncertainty ^d	Largest contributor
Dimming	Contributes to surface dimming	No direct effects on surface dimming
Acidification/fertilization	No ocean acidification/fertilization effects	Main contributor to ocean acidification and fertilization

- ^a UNEP and WMO (2011a) estimate narrower central range of +0.3 to +0.6 W m⁻².
- ^b Some adjustment to this value may be appropriate to account for the greater warming efficacy of BC deposited on snow and ice. UNEP and WMO (2011a) suggest a range of +0.05 to +0.25 W m⁻² for BC snow/ice albedo forcing (adjusted for efficacy).
- ^c Values are highly uncertain. The IPCC estimated that the cloud albedo effect of all aerosols combined was -0.7 W m⁻², but did not include other cloud effects and did not estimate the albedo effect of BC alone. UNEP and WMO (2011a) provided a central forcing estimate for all of the cloud impacts of BC of -0.4 to +0.4 W m⁻².
- ^d Based on the IPCC forcing estimates (shown in Figure 2-10), the central estimate of the UNEP assessment (UNEP and WMO, 2011), and this report’s assessed range of BC forcing (from Figure 2-11), it is likely that the net BC forcing will be less than that of CH₄ but it is possible that BC might be the second largest contributor to warming, depending on uncertainties in the direct and indirect contributions of BC to warming.
- ^e The lifetime of CO₂ is more complicated than for most other GHGs. The carbon in CO₂ cycles between the atmosphere, oceans, ecosystems, soil, and sediments. Carbon added to the carbon cycle is removed very slowly (over thousands of years) through processes such as weathering and calcium carbonate formation (Archer et al., 2009). However, even if it is not removed from the carbon cycle, carbon added to the atmosphere can also cycle to other media: approximately three quarters of the added carbon will, over a time scale of decades or centuries, move out of the atmosphere into the ecosystem or oceans.
- ^f Forster et al. (2007).

composition. For example, atmospheric aging can increase the size of a particle or alter its chemical composition in a way that makes it an efficient nucleus for cloud droplet formation, facilitating its removal by precipitation.

By contrast, GHGs have longer atmospheric lifetimes. This enables them to become well mixed in the atmosphere and to continue to absorb energy over many decades or centuries. Gases such as nitrous oxide (N₂O), CH₄, or hydrofluorocarbons (HFCs) have lifetimes that range from as short as a year for some of the HFCs to as long as 50,000 years for tetrafluoromethane (CF₄), a perfluorocarbon

(Forster et al., 2007).⁶ The carbon in CO₂ cycles between the atmosphere, oceans, ecosystems, soil, and sediments; therefore, CO₂ does not have a single defined lifetime. Computer models have indicated that about half of an emissions pulse of

⁶ Several months are required in order for a gas to mix throughout a hemisphere, and one to two years are required for a gas to become well-mixed globally. Some gases, such as CH₄, are well-mixed but are included in the category of “short-lived climate forcer” because a decadal lifetime has different implications for mitigation decisions than the lifetimes of a century or more for many other GHGs. On the other hand, ozone and water vapor have short lifetimes and, like black carbon and other aerosols, are not well-mixed in the atmosphere.

CO₂ will disappear within 30 years, 30% within a few centuries, and the last 20% may remain in the atmosphere for thousands of years (Denman et al., 2007).

BC's short atmospheric lifetime means that atmospheric concentrations are highest near significant emissions sources and during time periods and seasons of emissions releases. This **high spatial and temporal variability** affects BC's impacts on climate. BC is a regional pollutant. CO₂ and other GHGs with lifetimes longer than a year are global pollutants with relatively uniform concentrations around the globe. It is generally assumed that CO₂ and other well-mixed GHGs have essentially the same effect on climate regardless of the location or season of emissions. The same is not true for BC.

Geographic location and altitude are important determinants of the impact of BC on climate. Fine combustion particles including BC can be transported up to thousands of miles from sources. Particles have a greater effect on the net absorption of solar radiation by the atmosphere when they are emitted or transported over light-colored, reflective (i.e., high "albedo") surfaces such as ice, snow, and deserts. In the absence of PM, a high percentage of sunlight would reflect off these surfaces and return to space. Therefore, any absorption of either incoming or reflected light by PM above these surfaces is more likely to lead to warming than absorption of light by PM above darker surfaces. Even PM that is typically classified as reflecting can darken these bright surfaces and contribute to warming (Quinn et al., 2011). This mechanism explains why studies have found the effects of BC to be magnified in the Arctic and other alpine regions, as discussed in sections 2.6.4 and 2.6.5. In addition, the net radiative effects of BC can be sensitive to altitude. A modeling study by Ban-Weiss et al. (2011) suggests that while BC at low altitudes (where most BC is indeed located) warms the surface considerably, BC at stratospheric or upper-tropospheric altitudes may decrease surface temperature. In addition, as with particles suspended above a bright desert or glacier, particles suspended above bright cumulus clouds can absorb both incoming and outgoing solar radiation, increasing the net radiative effect of the light absorbing particle. When suspended between cloud layers or beneath a cloud, the particle may be shielded from incoming light, therefore lessening its potential radiative impact (Schulz et al., 2006).

Other key distinguishing features of BC include the wide **range of mechanisms** through which it influences climate and its association with other

adverse, non-climate related **public health and welfare effects**. In addition to the direct radiative forcing characteristic of both BC and GHGs, BC has significant interactions with clouds that can result in both warming and cooling effects. It can also cause melting and warming via deposition to snow and ice. BC and other particles are also directly associated with a host of other environmental effects, such as changes in precipitation patterns and surface dimming. All of these effects are discussed in greater detail later in this chapter. GHGs, on the other hand, influence climate mainly through direct radiative forcing effects. GHGs do not directly interact with clouds, snow and ice, though the warming of the atmosphere due to GHGs does influence cloud formation, snow melt, and many other climate properties. In addition, CO₂ has a fertilization effect on plants and an acidification effect in the ocean, and CH₄ emissions lead to increased ozone concentrations and changes in the lifetime of other atmospheric pollutants. Finally, as a constituent of PM_{2.5}, BC is directly linked to a range of public health impacts (see Chapter 3). This, too, distinguishes it from long-lived GHGs, which affect public health and welfare primarily via climate change effects.

An important implication of BC's strong absorptive capacity, coupled with its short atmospheric lifetime, is that when emissions of BC are reduced, atmospheric concentrations of BC will decrease immediately and the climate, in turn, will respond relatively quickly. The **potential for near-term climate responses** (within a decade) is one of the strongest drivers of the current scientific interest in BC. Mitigation efforts that reduce BC emissions can halt the effects of BC on temperature, snow and ice, and precipitation almost immediately. This means that reductions of BC may have an immediate and important benefit in slowing the near-term rate of climate change, especially for vulnerable regions such as the Arctic and the HKHT region. In contrast, when long-lived GHG emissions are reduced, the climate takes longer to respond because atmospheric GHG concentrations—the result of cumulative historic and present-day emissions—remain relatively constant for longer periods (see, for example, Figure 2-7). It is important to recognize, however, that the short atmospheric lifetime of BC also means that reductions in current BC emissions will have much less impact on temperature many

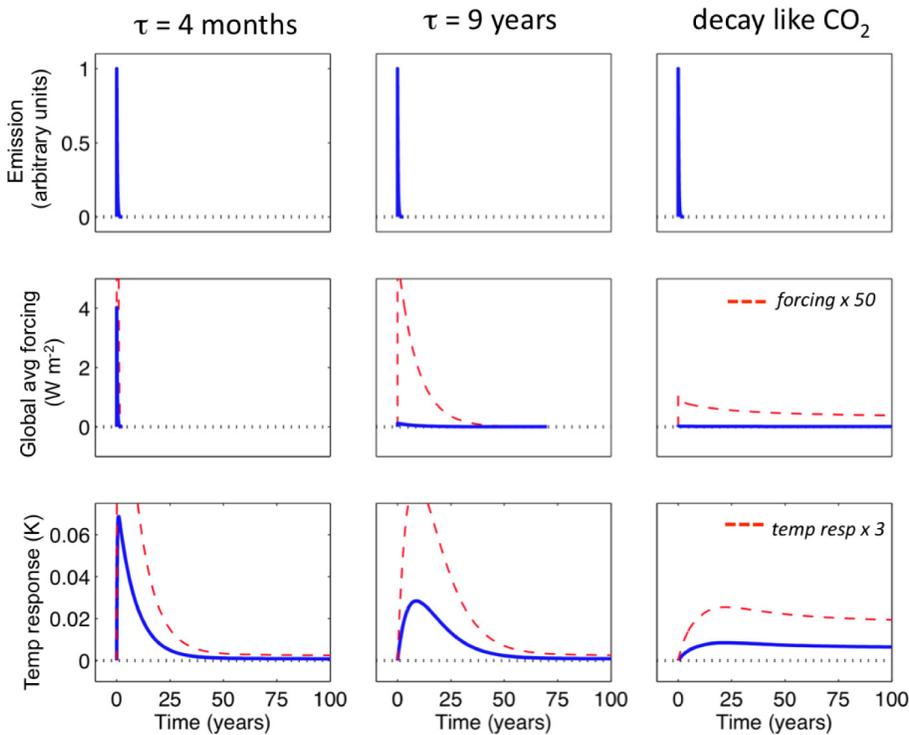


Figure 2-7. Climate Response to Emissions of Pollutants with Different Lifetimes. Figure illustrates forcing and temperature response to pulse emissions of three hypothetical substances with different atmospheric lifetimes (Column 1 illustrates short-lived species like BC, Column 2 illustrates species with medium atmospheric lifetime like methane, and Column 3 illustrates a long-lived species like CO₂). In the middle row, which illustrates global average forcing, the dashed red trace shows forcing value times 50 to show more detail in the time dependence. (Bond et al., 2011)

decades from now,⁷ which is mainly driven by past, current, and future emissions of persistent species like CO₂ combined with emissions of BC at that future date. Therefore, only sustained reductions in long-lived GHGs can avert long-term climate change.

In order to avoid (rather than delay) reaching temperature thresholds, long-lived GHG emissions reductions are necessary, but future BC reductions in conjunction with these GHG reductions can “shave the peak” off the temperature change and help avoid crossing these thresholds (possibly avoiding key non-linear climate impacts, often referred to as “tipping points”) (Lenton et al., 2009). Figure 2-8 demonstrates this visually: reductions in emissions of CH₄ and BC [the specific measures involved are described in more detail in the UNEP report (UNEP and WMO, 2011a)] lead to a near-term reduction in the rate of warming, but in the long run serve only to delay any given temperature change. While CO₂ reduction measures do little to slow near-term warming (in this case, in part because they are accompanied by simultaneous reductions of

cooling aerosols), they are necessary for long-term stabilization of the climate. Reducing short-lived forcers in conjunction with CO₂ enables both a reduction in near-term warming (due to near-term BC and CH₄ emissions reductions) and a decrease in peak warming (due to future BC and CH₄ reductions combined with CO₂ reductions over the entire time period).

These differences between BC and GHGs have significant implications for BC mitigation decisions. Specifically, the effectiveness of a given mitigation effort depends on the timing and location of the emissions; the atmospheric processes, transport, and deposition rates of the emissions from the specific sources; the underlying surface (e.g., ice and snow); and the presence of co-pollutant emissions. This constitutes a significant difference from long-lived GHGs where the precise timing and location of emissions (or emissions reductions) does not matter significantly with respect to the climate impact.

2.5 The Role of Co-Emitted Pollutants and Atmospheric Processing

As discussed in section 2.3, BC is never emitted into the atmosphere in isolation. Rather, it is always emitted as part of a mixture of particles and gases during the combustion process. The composition of this mixture can vary significantly, depending on combustion conditions and fuel type. BC is generally

⁷As shown in Figure 2-7, the temperature response to an immediate reduction in BC can last a couple of decades – due mainly to inertia in the climate because of how the ocean takes up heat. However, not included in this figure is the possibility that these temperature changes can have feedbacks which might persist for even longer – for example, higher temperatures reduce the rate at which CO₂ is mixed into the ocean, so a change in BC emissions today could change CO₂ concentrations, and therefore temperatures, for many decades.

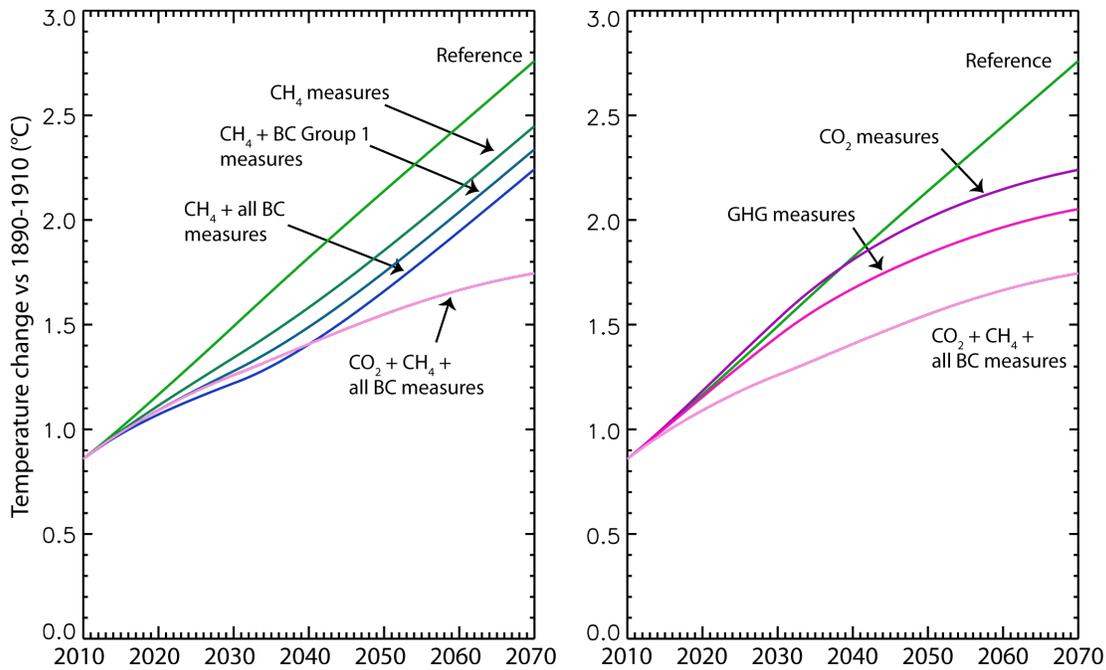


Figure 2-8. Projected Global Mean Temperatures under Various Scenarios Relative to the 1890-1910 Average. This figure is adapted from Figure 6.2 from the UNEP/WMO Assessment. The left panel shows a reference scenario, and then a sequence of emissions reduction scenarios in which first a set of methane reductions are implemented, then one set of BC measures, then a second set of BC measures, and then a set of CO₂ measures. In the right panel, the same reference scenario is shown, followed by a sequence of emissions reduction scenarios in which first the same set of CO₂ measures is applied, followed by some additional GHG reductions (mainly methane), followed by the full set of reductions from the left panel. More details about these measures can be found in the UNEP/WMO report. (Note that modeled effects of all measures include reductions of co-emissions, including reductions in cooling aerosols, which in the case of CO₂ measures resulted in a slight warming effect in the early years.) (UNEP and WMO, 2011a)

accompanied by OC, including BrC and other carbonaceous materials. In addition, an emissions plume may contain water, inorganic potassium and sodium salts, ammonium nitrate and sulfate, gaseous constituents (e.g., SO₂, NO_x and VOCs), various hazardous air pollutants (e.g., metals), and even soil particles.

The absorptive properties of an emissions plume from a specific source will depend on all of the **co-emitted pollutants**, and on how these constituents interact with one another and other atmospheric constituents in the atmosphere. As described above, BC is co-emitted with OC and/or sulfate, nitrate and gaseous constituents (SO₂, NO_x, and VOCs). Since OC and sulfate and nitrate particles generally exert a net cooling influence, these pollutants play an important role in determining the net absorptive capacity of the emissions plume. These other constituents, however, may be emitted in greater volume than BC, counteracting the warming influence of BC. Thus, estimating

the climate impact of BC quantitatively requires accounting for the impact of these co-emitted pollutants. Emissions from a single source can also vary over time. For example, the flaming phase of a wildfire produces much more BC than its smoldering phase. Also, when diesel trucks are under load, they produce more BC than during other parts of their driving cycle. Total **particle number** also impacts scattering and absorption: the more particles present in a portion of the atmosphere, the greater the probability that light rays will be scattered or absorbed by some of these particles.

Emissions from particular sources are often characterized in terms of their OC to BC ratio. Sources whose emissions mixtures are richer in BC relative to the amount of OC emitted (i.e., with lower OC:BC ratios) are more likely to contribute to climate warming; therefore, mitigation measures focusing on these sources are more likely to produce climate benefits. These ratios are useful in that they take the emissions mixture into account;

however, they rely on crude accounting methods and cannot provide precise measures of a particular source's climate impacts. A particular concern is the common presumption that all OC is cooling, when in fact some components (especially BrC) are light-absorbing and may contribute to the warming associated with an emissions mixture. (The use of OC:BC ratios is discussed further in Chapter 7.)

A fresh emissions plume contains particles of many different chemical compositions. Atmospheric scientists refer to an emissions plume with this kind of high inter-particle chemical variability as "**externally mixed**." The externally mixed plume, however, undergoes rapid chemical and physical transformations. The coagulation of particles, assuming no secondary particles form through atmospheric chemical reactions, reduces the overall particle number. This process, combined with thermodynamically-driven mixing processes such as water condensation and the redistribution of semi-volatile PM compounds, reduces the differences in chemical composition among the individual particles in the plume. Over time, the chemical composition of the particles within a given plume approaches uniformity. A plume is said to be

"**internally mixed**" when it is near this theoretical end-point. *In situ* measurements indicate that emissions become internally mixed within a few hours (Moffet and Prather, 2009b).

BC is an insoluble material, thus a well-mixed or "aged" emissions plume will contain particles where BC appears as an inclusion in an otherwise liquid particle. These particles are often described as "coated" BC particles. As described in section 2.3, common coatings include sulfuric acid, water, and transparent OC. The degree of mixing, or more specifically, the fraction of BC particles that are coated, influences the absorptive properties of the particle. Internal mixtures of particles that include BC have been observed to absorb light more strongly than pure BC alone, by 30 to 100%. Whether BC is modeled as an externally or internally mixed particle can have a large effect on resulting estimates of radiative forcing (see section 2.6.1.2).

Emissions plumes from different sources interact with each other as well as with the surrounding atmosphere. As a combustion emissions plume rises into the atmosphere, it is diluted by ambient air (see Figure 2-9). The open atmosphere contains

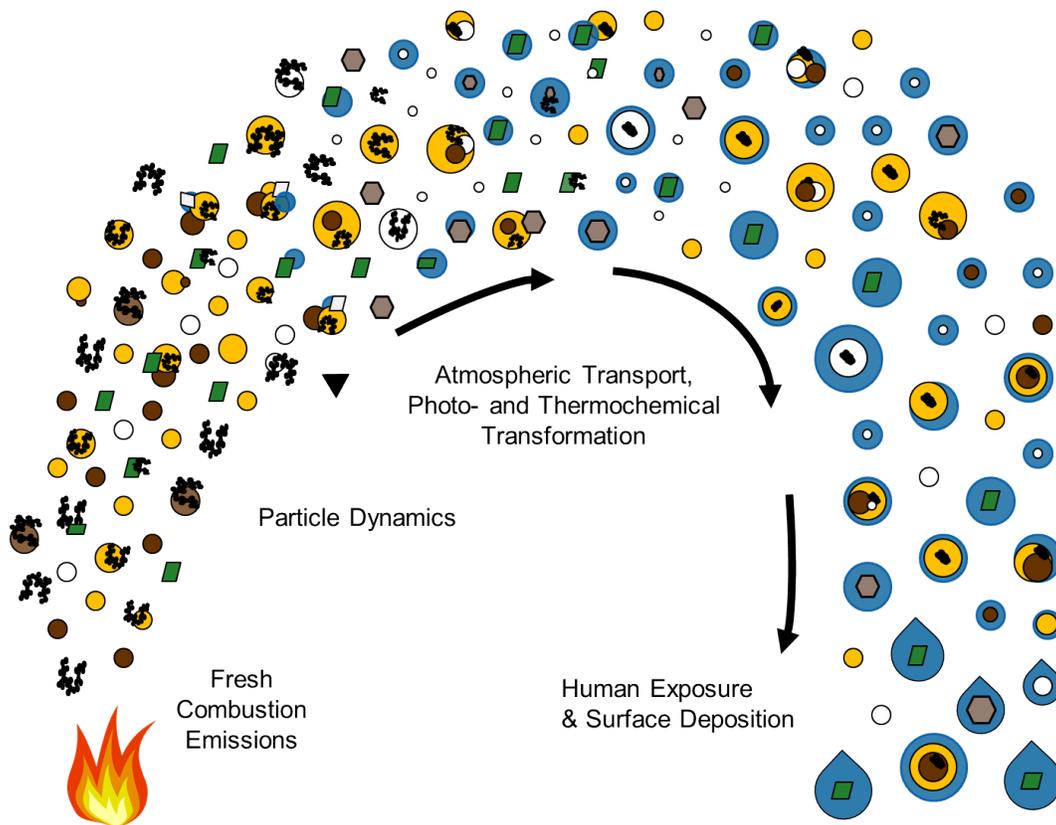
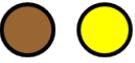


Figure 2-9. Particle Transformation in the Atmosphere, from Point of Emission to Deposition. A variety of physical and chemical processes contribute to changing the light-absorption capacity of a fresh plume. (Source: U.S. EPA)

a number of reactive gases and particles originating from a wide variety of anthropogenic and natural sources. Physical and chemical changes resulting from coagulation, condensation, and other photochemical and atmospheric processes can alter the climate-forcing impact of a given emissions plume (Lauer and Hendricks, 2006) (see Table 2-2). These mixing and aging processes are complex. However, excluding them from models of the climate impacts of BC may yield incomplete or erroneous estimates. For example, coating of a BC particle by

a clear (light-scattering) shell has been shown to enhance light absorption because the shell acts as a lens that directs more light toward the core (Ackerman and Toon, 1981; Jacobson, 2000; Lack and Cappa, 2010; Jacobson, 2001). Other authors have also found that light-absorption by BC is enhanced when BC particles are coated by sulfate or other light scattering materials (Shiraiwa et al., 2009; Sato et al., 2003; Moffet and Prather, 2009a; Bond et al., 2006a). Other atmospheric processes, however, such as further chemical processing or

Table 2-2. Examples of Particle Types and Mixtures Present in Combustion Plumes. The size, shape, and chemical composition of a particle or particle mixture determine its radiative properties.

Particle	Type	Radiative Properties
	Black carbon ^a	Absorbing (all solar wavelengths)
	Brown (or yellow) carbon ^b	Absorbing (UV and some visible)
	Non-absorbing carbon ^b	Scattering
	Nitrate ^c	Scattering
	Sulfate ^c	Scattering
	Black carbon coated with brown or non-absorbing carbon ^d	Absorbing (enhanced by partial internal reflection of solar radiation); fractionally scattering
	Black carbon associated with sulfate or nitrate ^e	Absorbing plus some scattering
	Cloud and fog droplets ^f	Scattering
	Complex of several particles ^e	Absorbing and scattering
	Mixed particle (cloud processed) ^g	Absorbing (enhanced by partial internal reflection of solar radiation); fractionally scattering

- ^a Fresh BC is produced primarily during flaming combustion.
- ^b Particles condense within a fresh combustion plume from pyrolytic BrC and yellow OC. Oxidation of anthropogenic and biogenic VOCs produces non-light absorbing carbon particles, and may also produce BrC and yellow carbon PM.
- ^c Emitted directly as a byproduct of combustion, or formed through the oxidation of SO_x or NO_x.
- ^d In the exhaust gases of solid fuel fires, low volatility BrC and other organic compounds can condense on BC particles. In the ambient atmosphere, low volatility organics produced by oxidation of VOCs can also condense on BC.
- ^e Forms when high particle concentrations lead to the coagulation of multiple particles.
- ^f Forms by condensation of water vapor onto acidic organic (carbon-based) and inorganic particles.
- ^g Forms when complex particles undergo the humidification and drying cycles characteristic of cloud formation and evaporation.

particle growth through coagulation, may off-set this enhancement. Observations by Chan et al. (2011) at a rural site in Ontario, where BC particles were assumed to be coated, did not show enhanced light absorption. Furthermore, coated particles are more easily removed by cloud droplets and precipitation, decreasing their atmospheric lifetime (Stier et al., 2006).

2.6 Global and Regional Climate Effects of Black Carbon

BC affects climate through both direct and indirect mechanisms. The most extensively studied of these mechanisms is radiative forcing, which is directly linked to temperature change. Radiative forcing is a measure of how a pollutant affects the balance between incoming solar radiation and exiting infrared radiation, generally calculated as a change relative to preindustrial conditions defined at 1750. A pollutant that increases the amount of energy in the Earth's climate system is said to exert "positive radiative forcing," which leads to warming.⁸ In contrast, a pollutant that exerts negative radiative forcing reduces the amount of energy in the Earth's system and leads to cooling. The net radiative impact of a pollutant since preindustrial times can be averaged over the Earth's surface and is expressed in Watts per square meter ($W m^{-2}$). Global average radiative forcing is a useful index because in general it is related linearly to the global mean temperature at the surface (Forster et al., 2007) and is approximately additive across pollutants. Radiative forcing also provides a consistent measure for comparing the effects of past and projected future emissions. As a result, it has become a standard measure for organizations like the IPCC and the U.S. Global Change Research Program (National Research Council of the National Academies, 2005).

In addition to radiative forcing, BC is associated with other effects including surface dimming and changes in precipitation patterns. While not directly linked to net global temperature change, these effects also have important global and regional climate implications. Each of these effects is discussed in greater detail later in this section.

This section mainly addresses the climate impacts of BC, GHGs, and other substances based on the radiative forcing resulting from the change in concentrations of these substances since

⁸ In general, radiative forcing in this document refers to "top-of-the-atmosphere" radiative forcing unless otherwise specified. Measuring at the top of the atmosphere (TOA) (in this context, between the troposphere and the stratosphere) is the best location for determining net energy balance.

Radiative Forcing

Radiative forcing: The change in the energy balance between incoming solar radiation and exiting infrared radiation, typically measured in watts per square meter ($W m^{-2}$), due to a change in concentration (generally the change since preindustrial conditions in 1750). Positive radiative forcing tends to warm the surface of the Earth, while negative forcing generally leads to cooling.

preindustrial times. There are, however, other ways to address climate impacts, and the decision of which approach to use depends on at least two key issues.

The first issue is that an analysis based on changes in concentrations since preindustrial times is in some ways a historical measure. An alternative approach would be to analyze the effect of current day emissions on radiative forcing into the future [e.g., Figure 2-19 in this report, Figure 2.22 in Forster et al. (2007), or the analysis by Shindell et al. (2009)]. The benefit of such an approach is that an emissions-based analysis is more policy-relevant, because policies directly control emissions and not concentrations. There are several disadvantages to using this approach. First of all, fewer studies have used this approach. Second, the approach requires choosing a time frame of integration (see section 2.7 on metrics for more discussion of time frames). Third, there are increased uncertainties because the results depend on decisions regarding background concentrations and are more sensitive to model factors such as atmospheric chemistry and carbon cycles. In contrast, analyses based on existing concentrations are less uncertain because for the most part they depend on measured concentrations. In general, this is more important for the long-lived GHGs than for short-lived substances such as BC, and thus becomes an issue for purposes of comparisons.

The second issue in addressing climate impacts is whether to consider radiative forcing as an endpoint, or whether to actually calculate impacts on temperatures. The advantage to examining temperatures is that they are more immediately relevant to human experience than the more abstract "radiative forcing". Also, though radiative forcing is approximately additive in most cases, there are some exceptions: as discussed in section 2.6.1.4, temperatures are actually much more sensitive to snow albedo forcing than to other forcings.

However, there are two disadvantages to the temperature approach. The first is that, again, fewer studies use this approach, in part because the models used to calculate temperature are more computationally intensive than those that calculate radiative forcing. The second is that the relationship between radiative forcing and temperature, while generally linear, is very model dependent. Every model has an inherent “climate sensitivity”, and therefore when comparing the impacts of BC from one model to the next, differences in this climate sensitivity can make it difficult to understand if differences between the models are due to BC physics or to the general model response to any forcing.

2.6.1 Global and Regional Radiative Forcing Effects of BC: Overview

This section provides an overview of the different effects of BC on radiative forcing based on the best estimates in the current literature. These effects include **direct forcing** (direct absorption of solar or terrestrial radiation), **snow/ice albedo forcing** (forcing that results from the darkening of snow and ice), and **indirect forcing** (a range of forcing effects resulting from impacts on clouds, including changes in cloud lifetime, reflectivity, and composition). Section 2.6.1.1 describes the overall net effects of BC on radiative forcing, when the effects of these different types of forcing are accounted for. The subsequent sections present more detailed information regarding the state of knowledge regarding the specific effects of BC on direct forcing, snow/ice albedo forcing, and indirect forcing at the global and regional scales. Each section provides a summary of the findings of recent studies, explanations of the differences among estimates, and characterizations of key remaining uncertainties.

In evaluating estimates of the effects of BC on radiative forcing, it is necessary to consider several caveats. For example, it is important to differentiate among estimates with respect to the baseline time period used to define the radiative forcing estimates. The radiative forcing estimates (and other climate effects) are often expressed as a comparison to a given historical level rather than with respect to present day or in terms of the anthropogenic influence compared to total forcing. However, these assumptions are not always stated clearly. Also, it is important to differentiate with respect to the types of BC emissions included. For example, many studies exclude open biomass burning. The inclusion or exclusion of BC from wildfires and other sources of open biomass burning will affect the estimates of net BC effects. In addition, because some studies evaluate the climate effects of BC as it co-occurs with

other aerosol chemical species, such as OC, sulfates and nitrates, while others do not, it is important to distinguish studies where BC is estimated individually from studies where BC is estimated as part of an aerosol mixture. In the following sections, it is indicated whether the radiative forcing estimates include co-occurring OC and other species and how these other pollutants influence estimates of BC’s global and regional climate impacts, when possible.

2.6.1.1 Net Forcing

As is described in more detail in the following sections, the different kinds of forcing involve different mechanisms of action and can have offsetting climate effects. For example, direct effects are associated with positive forcing, while most (but not all) indirect effects are thought to result in negative radiative forcing. This section provides an overview of estimates of the direction and magnitude of the net effect of BC on radiative forcing when the direct forcing, snow/ice albedo forcing, and indirect forcing effects of BC are summed, and identifies key factors that contribute to variability in these estimates

There is a range of quantitative estimates in the literature for global average radiative forcing due to BC. Most studies indicate that due to the direct and snow/ice albedo effects, the net effect of BC on climate is likely to be warming. However, because of the large remaining uncertainties regarding interactions of BC with clouds, it is difficult to establish quantitative bounds for estimating global net impacts of BC, or even to completely rule out the possibility of a net negative effect.

The most widely utilized estimates of forcings for GHGs come from the IPCC’s Fourth Assessment Report, which was issued in 2007. The IPCC also estimated forcing due to BC based on a review of the scientific studies available at the time, though estimates from more recent studies surveyed in this EPA report differ somewhat as detailed further below. The IPCC estimated a direct radiative forcing of $+0.34 \text{ W m}^{-2}$ for BC, making BC third only to CO_2 and CH_4 . In addition, the IPCC estimated BC’s snow/ice albedo forcing to be $+0.1 \text{ W m}^{-2}$ (see Figure 2-10). Other aerosols were generally shown to have a cooling influence on climate. The IPCC estimates of negative direct radiative forcing due to OC and sulfates are also shown in Figure 2-10. Indirect effects for *all* aerosols, including BC, are also estimated to result in net negative forcing due to increased reflectivity of clouds (“cloud albedo effect”). The IPCC did not provide quantitative estimates of the effect of aerosols on other properties of clouds (such as lifetime, stability, etc.),

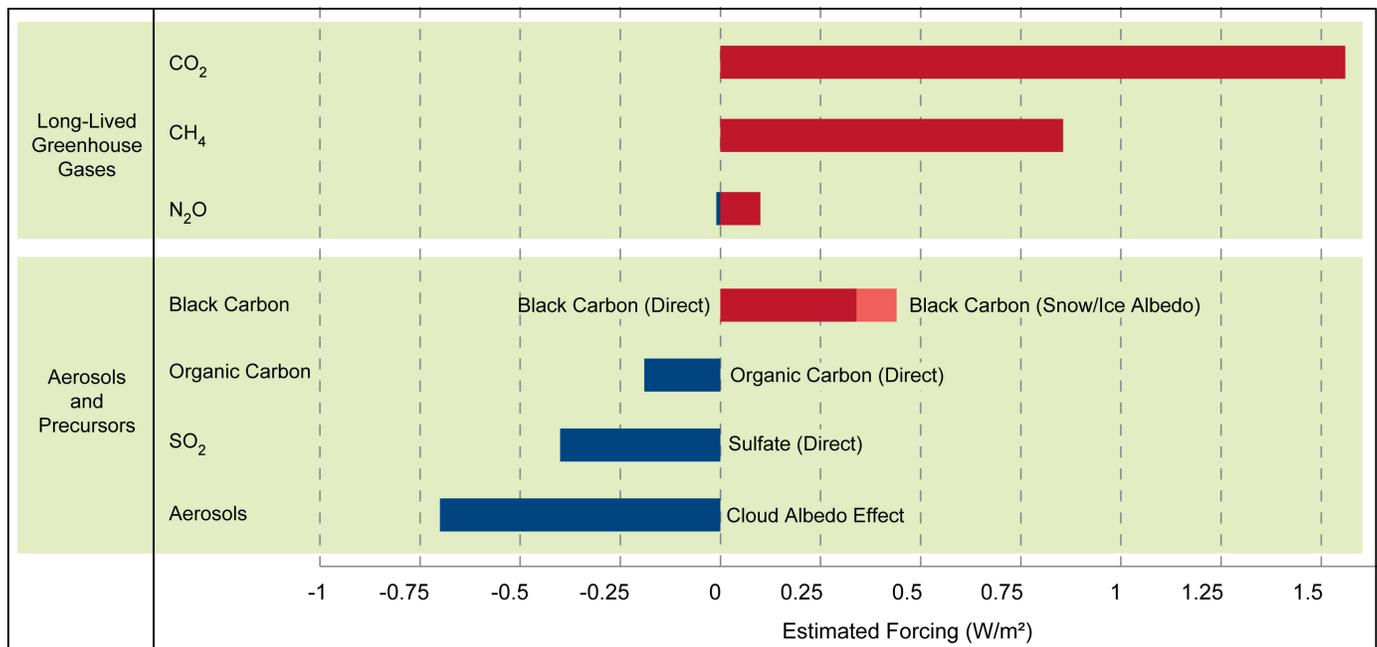


Figure 2-10. Components of Global Average Radiative Forcing for Emissions of Principal Gases, Aerosols, and Aerosol Precursors, based on IPCC estimates. Values represent global average radiative forcing in 2005 due to emissions and changes since 1750. Total radiative forcing for CH₄ includes the effects of historical CH₄ emissions on levels of tropospheric O₃ and stratospheric H₂O, and the CO₂ oxidation product of CH₄ from fossil sources. Similarly, total radiative forcing for N₂O includes the effect of historical N₂O emissions on levels of stratospheric O₃. The IPCC does not report an overall uncertainty for the net contribution to forcing of individual GHG emissions. However, based on the uncertainties provided for the individual components of these contributions, the uncertainty in forcing from CO₂ and N₂O emissions is extrapolated as being approximately 10% and approximately 20% from CH₄ emissions. Uncertainty in direct forcing is ±0.25 W m⁻² for BC and ±0.20 W m⁻² for both OC and SO₂. The range of forcing for the cloud albedo effect is -1.8 to -0.3 W m⁻². (Adapted from Figure 2.21 of Forster et al., 2007)

or quantify the indirect effects of individual aerosol species (such as BC) separately. As a result, there is substantial uncertainty in the IPCC’s estimates of net forcing for BC.

In addition to the estimates compiled by the IPCC (2007), many other studies have attempted to estimate the global average radiative forcing attributable to BC. An examination of the results of these studies, as summarized in Figure 2-11, indicates that the direct effect and the snow/ice albedo effect of BC are positive, though the magnitude of these effects is uncertain. The figure shows the range of central estimates from the included studies (solid box) as well as the highest and lowest uncertainty estimates from those studies (error bars) for both the direct effect and the snow and ice albedo effect. As discussed further below, a number of studies have estimated BC’s direct radiative forcing to be higher than the IPCC estimate (Sato et al., 2003; Ramanathan and Carmichael, 2008).

The biggest source of uncertainty about the net forcing effect of BC is the magnitude of the cloud effects of BC. (Cloud effects are discussed in detail in section 2.6.1.3.) The limited number of studies in the literature allow for statements on the direction (e.g., warming or cooling) of some of these forcings, but not their magnitude, as shown in Figure 2-11. The impact on cloud lifetime and albedo is likely cooling. The interactions with mixed-phase and ice clouds are likely to be warming. Semi-direct effects are uncertain, and existing studies differ on the definition and net influence of the effect. The cloud absorption effect is positive.

In light of the large remaining uncertainties about the magnitude (and in some cases the sign) of the different forcing effects of BC, particularly with regard to the cloud interaction effects, this EPA report does not assign a range to the magnitude of the net effect beyond noting that it is very likely to be positive (however, a net negative effect cannot be excluded). As indicated in Figure 2-11, the

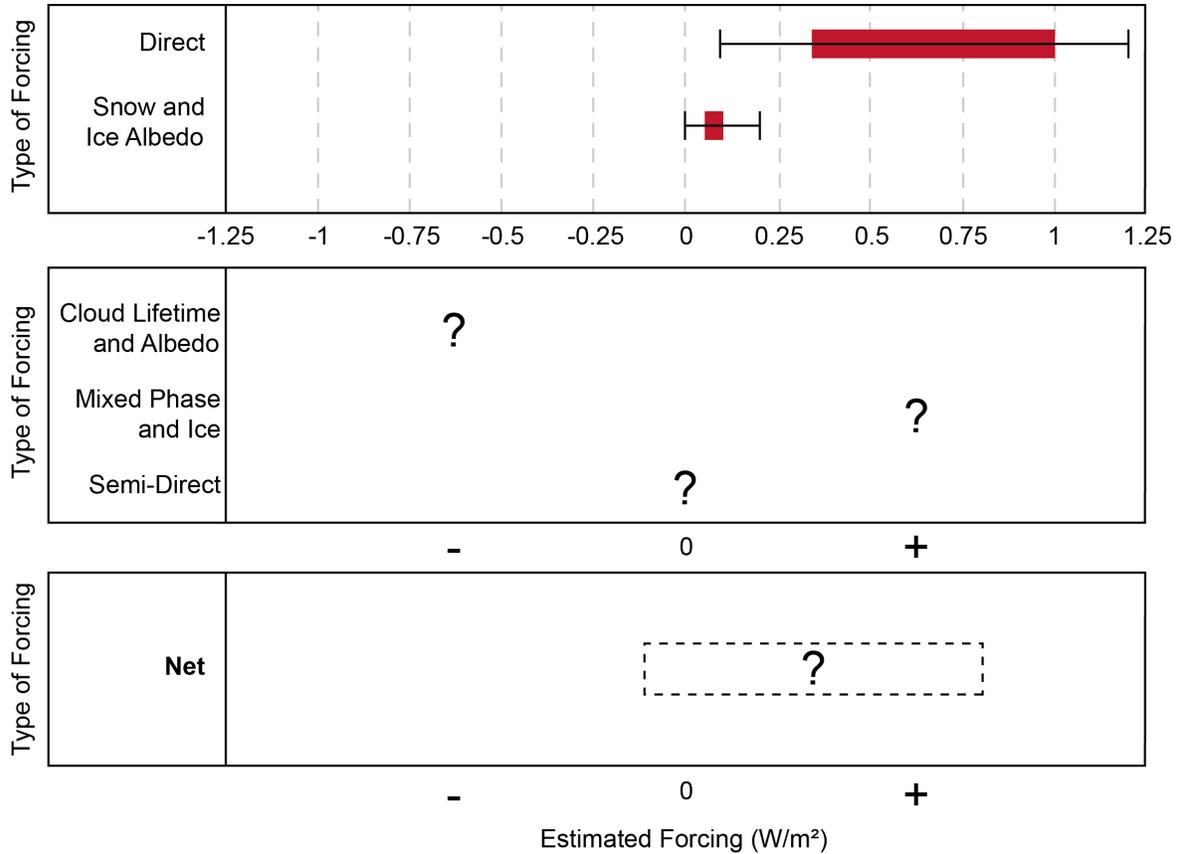


Figure 2-11. Estimates of Radiative Forcing from BC Emissions Only. The boxes indicate ranges of central estimates from the papers identified in this report, with error bars indicating the highest and lowest uncertainty estimates from those papers. Estimates are based on a synthesis of results from eleven studies that considered the direct forcing effects of BC emissions and six studies that considered reduction in snow and ice albedo from BC emissions. The range for the snow and ice albedo bar does not include the effects of the higher efficacy of the snow albedo effect on temperature change (forcing efficacy is described in section 2.6.1.4). The studies of indirect and semi-direct radiative forcing effects due to BC emissions are not sufficiently comparable in scope and approach to combine the estimates. As a result, only the likely direction of forcing is presented. Two other warming effects, the cloud absorption effect and the water vapor effect, have recently been discussed in the literature but are not included here. (Source: U.S. EPA)

estimates of direct and snow/ice albedo forcing are likely to be positive, but additional work is needed to determine the extent to which these positive forcing effects are offset by indirect effects, semi-direct effects, and other effects on clouds.

The recent UNEP/WMO assessment (UNEP and WMO, 2011a) evaluated a number of recent studies to investigate the net effect of BC on climate. Based on estimates of BC forcing due to direct, indirect, and snow/ice albedo effects, UNEP/WMO estimated that the global average net forcing was likely to be positive and in the range of 0 to 1 W m⁻², with a central value of 0.6 W m⁻² (this estimate included an enhanced efficacy factor for the snow/ice albedo effect: forcing efficacy is described in section

2.6.1.4). The UNEP/WMO assessment estimates for the different forcing effects of BC are shown in Table 2-3. In selecting a central estimate for net BC forcing, the authors of the UNEP/WMO assessment noted the very strong negative forcing due to total aerosols,⁹ and the current lack of quantitative

⁹The IPCC estimated the net forcing due to all aerosols as -1.2 W m⁻² (including a direct forcing effect of -0.5 W m⁻² (-0.9 to -0.10 W m⁻²) and a cloud albedo effect of -0.7 W m⁻² (-1.8 to -0.3 W m⁻²) (IPCC, 2007). However, these estimates are assigned a low level of scientific understanding by IPCC. More recent estimates in the literature continue to span a large range: Bauer and Menon (2012) estimate the net aerosol effect as -0.6 W m⁻² (including direct -0.5 W m⁻² and indirect -0.1 W m⁻²), while Hansen et al. (2011) argue the effect is more strongly negative (-1.6 W m⁻² ±0.3)

Table 2-3. Summary of UNEP/WMO Assessment Estimates of Radiative Forcing Effects of BC. (Adapted from UNEP and WMO, 2011a, Table 3-1)

Forcing	Range (W/m ²)	Central Values (W/m ²)	Mean Value (W/m ²)
Direct	0.2 to 0.9	0.3 to 0.6	0.45
Semi-direct and indirect	-0.6 to 0.5	-0.4 to 0.4	0.0
Deposition	(0.01 to 0.08) corrected for efficacy (x5) 0.05 to 0.4	(0.01 to 0.05) 0.05 to 0.25	0.03 0.15
Total BC	-0.3 to 1.8	0.0 to 1.3 0.0 to 1.0 with constraints on total aerosol forcing	0.5 0.6 including efficacy of deposition

estimates for all processes. In light of these constraints, they argued that net effective BC forcing is unlikely to exceed 1 W m⁻².

There remains a strong need for further research to provide better estimates of the forcing effects of BC and to reduce remaining uncertainties. Work in progress by a consortium of researchers under IGAC/SPARC (currently being prepared for submission to an academic journal with expected publication later in 2012) will likely provide more definitive quantitative bounds on the BC cloud interaction effects, and the net effects overall.

2.6.1.1.1 Factors that Contribute to Variability in Estimates

There are a number of factors that may contribute to the lack of consensus among modeled estimates of net global average radiative forcing from BC. Koch et al. (2009) attributed the range of estimates to differences in the aerosol microphysical calculations in the models (i.e., different estimates of how much solar radiation each unit of BC absorbs). The authors also pointed out key differences in models, such as the assumed values of various physical properties, and differences in the representation of vertical transport and cloud effects.

Variability in the estimates may also arise due to differences in experimental design and how the values are reported. Radiative forcing is commonly measured and reported as top-of-the-atmosphere (TOA) radiative forcing which captures all variations in energy over the entire atmosphere. This is appropriate for the well-mixed, long-lived GHGs, but perhaps not for BC, which exhibits high spatial variability. For example, the vertical distribution of BC in the atmospheric column and interactions with clouds lead to inputs of energy at different altitudes compared to the input of energy due to GHGs (see Ramanathan et al., 2001, and references therein).

Climate effects are also sensitive to the location of the BC emissions. For example, Arctic sea ice melting may be accelerated by BC emissions from northern latitudes, as discussed later in this chapter. Finally, radiative forcing metrics that focus on specific species do not generally capture co-pollutant interactions, which are very important for BC.

2.6.1.1.2 Regional Dynamics

Studies focusing on global average radiative forcing may overlook key regional dynamics associated with BC as a spatially heterogeneous pollutant. Many studies have found that BC's regional climate impacts are more pronounced than the contributions of BC to global average temperature change. In addition, certain regions of the world are more sensitive to or more likely to be affected by BC forcing, either due to transport and deposition (e.g., the Arctic) or high levels of aerosol pollution in the region (e.g., Asia). Global average radiative forcings for BC hide much of the regional variability in the concentrations and impacts. Note, however, that regional variability of BC forcing may exaggerate the regional variability of impacts, as temperature impacts usually occur over a larger area and longer time period than the forcing effects.

2.6.1.2 Direct Forcing

The direct effect of BC is to absorb solar radiation. As mentioned above and as shown in Figure 2-12, the IPCC (2007) estimated the global average radiative forcing of BC from all sources at +0.34 (±0.25) W m⁻². A subset of this forcing due to BC from fossil fuel combustion (mainly coal, petroleum and gas fuels) was estimated to be +0.2 (±0.15) W m⁻². Most studies published since the IPCC report have reported higher direct forcing values. The UNEP/WMO assessment (UNEP and WMO, 2011a) estimated that direct forcing of BC is most likely to be within the range 0.3 to 0.6 W m⁻². Additional work

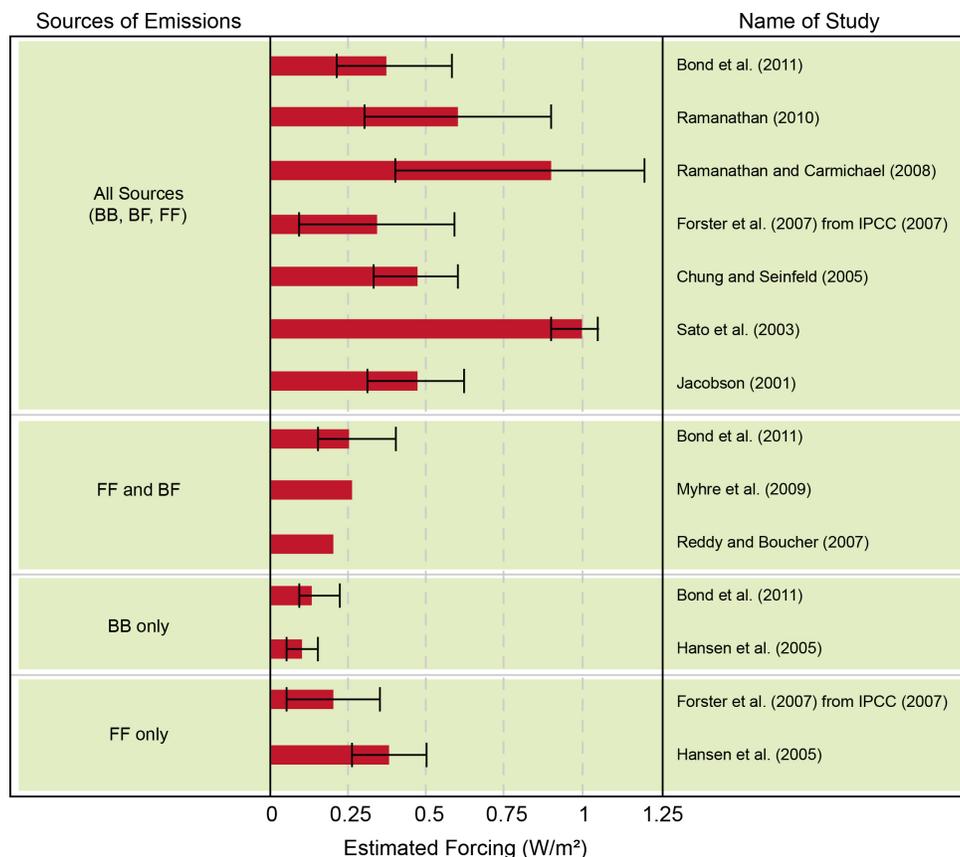


Figure 2-12. Estimates of Direct Radiative Forcing from BC Emissions Only. These values represent the range of estimates in the peer reviewed literature; however, they are not all directly comparable. Some are based on different estimates of BC emissions, include different sectors, and present the forcing with respect to different baseline time periods (e.g., 1750, post-industrial, present day). Note: BB = open biomass burning; BF = biofuels; and FF = fossil fuels. (Source: U.S. EPA)

is underway to try to develop a new central estimate for these direct impacts (IGAC/SPARC, forthcoming).

Assumptions about mixing state (e.g., internal/external) are critical to the results. As noted in section 2.5, studies that have incorporated internal mixing into the calculations of direct radiative forcing for BC yield higher forcing than those that do not, and these models are considered to be more realistic. Simulations by Jacobson (2001) found that accounting for internal mixing of BC in aerosols increases the estimated absorption and warming by BC by a factor of two. Koch et al. (2009) accounted for this underestimation of absorption by BC in older models by doubling the ensemble average from a 17-model intercomparison project (Schulz et al., 2006), resulting in a global average BC direct radiative forcing of roughly +0.5 W m⁻². Bond et al. (2011) combined forcing results from 12 models to use the best estimates for mixing and transport in those models. Based on this analysis, and using the same emissions estimates used by the models assessed in the IPCC reports, Bond et al. (2011) found a total forcing of +0.40 W m⁻², or 18% higher than the IPCC estimate, which they attributed to the fact that the IPCC estimate includes some models that do not include enhanced absorption due to internal mixing. Note that Bond et al. differentiate

“anthropogenic emissions” (post-1750, including open burning) and total BC emissions, calculating the total forcing from the latter to be +0.47 W m⁻².¹⁰

In addition to mixing state, other factors that can lead to variation in modeled radiative forcing effects of BC include particle size and removal (e.g., by precipitation). Vignati et al. (2010) describe how a modeled reduction in wet removal of BC from the atmosphere by 30% results in a 10% increase in BC’s atmospheric lifetime. In some cases, mainly in work based on observational constraints from the Aerosol Robotic Network (AERONET) ground-based sunphotometer network, much higher values have been reported. Sato et al. (2003) inferred a forcing of 1 W m⁻² based on these observational constraints. Chung et al. (2005) and Ramanathan and Carmichael (2008) combined the AERONET results with satellite data and report an estimated global average radiative forcing for BC of +0.9 W m⁻², with a range of +0.4 to +1.2 W m⁻². While most recent studies find global forcing higher than the IPCC, a discrepancy remains between the very high observationally constrained results and model results (even those

¹⁰ Ramanathan (2010) suggests that Bond et al. (2011) underestimate the total forcing, and suggests a central value of +0.55 with a 90% confidence interval of +0.2 to +0.9 W m⁻².

that include internal mixing and therefore produce higher values). Bond et al. (2011) hypothesized that the higher forcing in the observationally constrained results could result from higher emissions than in the model work. The exact cause of these differences, however, has not been isolated.

Compared to global radiative forcing, fewer studies have reported regional direct radiative forcing by BC. Studies such as Bond et al. (2011) show the geographic distribution of direct forcing from all sources of BC emissions (Figure 2-13). They found the largest forcing over South and East Asia and parts of Africa. Myhre et al. (2009), who considered only fossil fuel and biofuel BC, also found the largest forcing over South and East Asia. Other work such as Chung and Seinfeld (2005) showed similar patterns with higher forcing in Central and South America (the Amazon basin) and sub-Saharan Africa due to the inclusion of biomass burning emissions. Chung and Seinfeld (2005) report a range of +0.52 to +0.93 $W m^{-2}$ for externally and internally mixed BC respectively, averaged over the Northern Hemisphere. Their earlier work also suggests a strong seasonal cycle which peaks in May at +1.4 $W m^{-2}$ (Chung and Seinfeld, 2002). For the Southern Hemisphere, Chung and Seinfeld (2005) estimate a range of +0.15 to +0.23 $W m^{-2}$. Reddy and Boucher (2007) calculated the influence of regional BC emissions on the global average radiative forcing.

The largest contribution to global TOA BC radiative forcing came from East Asia (+0.08 $W m^{-2}$). The global average forcing due to North American BC emissions in this study was +0.02 $W m^{-2}$.

2.6.1.3 Cloud-related Forcings

The net effect of particles on climate via impacts on clouds is highly uncertain (IPCC, 2007). There are several different kinds of cloud effects that are important for radiative purposes, as summarized in Table 2-4. These cloud effects contribute to changes in the radiative balance of the atmosphere, and also influence climatic factors such as precipitation and dimming (section 2.6.3).

Since cloud droplets are formed when water vapor condenses onto a particle, many types of particles can affect the formation and microphysics of clouds. Emissions of aerosols into the atmosphere increases the number of particles on which cloud droplets can form, resulting in more and smaller cloud droplets. These additional cloud droplets produce brighter, more reflective clouds (Twomey, 1977). This generally results in surface cooling by preventing sunlight from reaching below the cloud to the Earth's surface (see also section 2.6.3.1 on surface dimming). This increase in reflectivity of the clouds has been termed the "the first indirect effect" or the "cloud albedo effect". In addition, the smaller

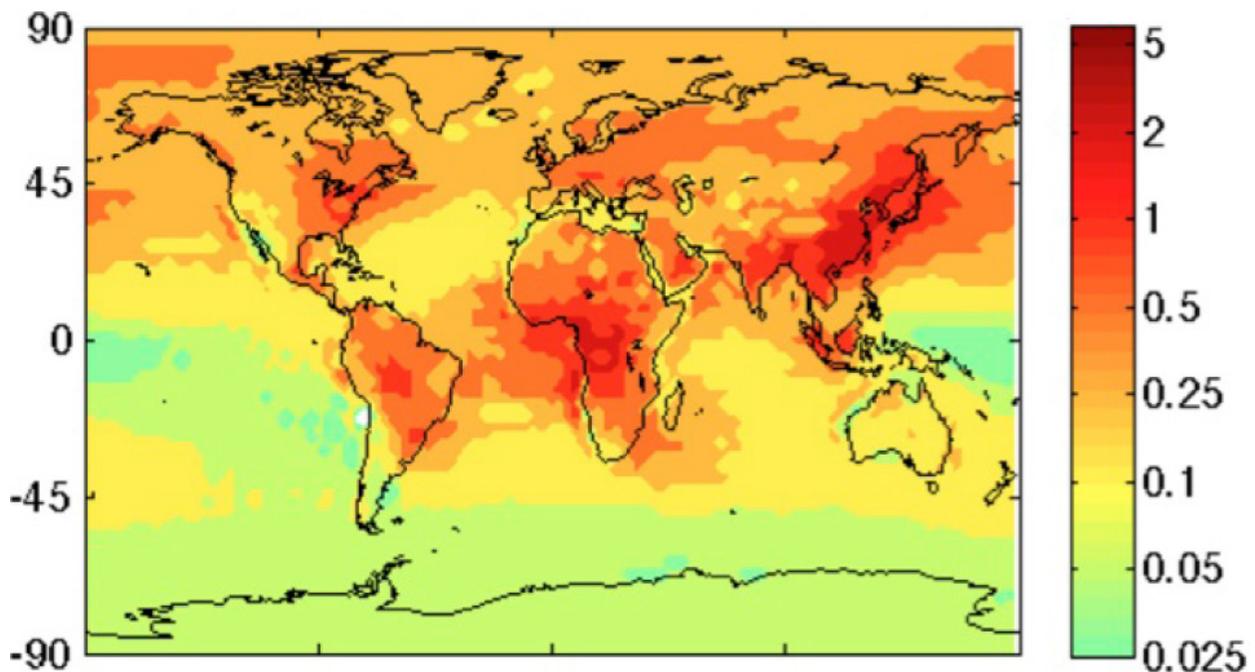


Figure 2-13. Direct Radiative Forcing ($W m^{-2}$) of BC from All Sources, simulated with the Community Atmosphere Model. (Bond et al., 2011)

Table 2-4. Overview of the Different Aerosol Cloud Effects. This summary applies to all aerosols and is not BC-specific, though BC is expected to participate to some extent in all of these mechanisms. Net effect refers to top-of-the-atmosphere radiative forcing. Scientific understanding is based on IPCC terminology. (Adapted from IPCC, Denman et al., 2007, Table 7.10a)

Effect	Cloud Types Affected	Process	Net Effect	Potential Magnitude	Scientific Understanding
Cloud albedo effect (first indirect effect)	All clouds	Smaller cloud particles reflect more solar radiation	Cooling	Medium	Low
Cloud lifetime effect (second indirect effect)	All clouds	Smaller cloud particles decrease precipitation so increase lifetime	Cooling	Medium	Very low
Semi-direct effect	All clouds	Absorption of solar radiation by absorbing aerosols changes atmospheric stability and cloud formation	Cooling or warming	Small	Very low
Glaciation indirect effect	Mixed-phase clouds	An increase in ice nuclei due to some aerosols increases precipitation	Warming	Medium	Very low
Thermodynamic effect	Mixed-phase clouds	Smaller cloud droplets delay freezing causing super-cooled clouds to extend to colder temperatures	Cooling or warming	Medium	Very low

cloud droplets are less likely to aggregate sufficiently to form rain drops, which changes precipitation patterns and increases cloud lifetime (Albrecht, 1989) (see also section 2.6.3.2 on precipitation impacts). This has been called the “second indirect effect” or the “**cloud lifetime effect**”. In general, the cloud albedo effect and the cloud lifetime effect are estimated to lead to cooling. However, Ramanathan (2010) asserts that the empirical evidence shows a positive forcing (warming) over land regions.

The magnitude and sign of the radiative effects depend on whether the clouds are composed of liquid droplets, ice particles, or a mix of ice and liquid droplets, and on the composition of the aerosol particles. In certain kinds of “mixed-phase clouds” (clouds with both ice and water), smaller droplets cause a delay in the freezing of the droplets, changing the characteristics of the cloud; however, the IPCC was not able to determine whether this “**thermodynamic effect**” would result in overall warming or cooling (Denman et al., 2007).

The “**semi-direct effect**” is specific to BC and other absorbing aerosols, while the “**glaciation indirect effect**” appears to be important for aerosols, such as BC and mineral dust, which are not especially hydrophilic. The semi-direct effect refers to the heating of the troposphere by absorbing aerosols, affecting the relative humidity and stability of the troposphere, which in turn affects cloud formation and lifetime (IPCC, 2007; Ackerman et al., 2000).

Older literature refers to the semi-direct effect as cloud burn-off (i.e., a decrease in cloud formation) from BC within the cloud layer. The definition was extended to include all effects on cloud formation and lifetime as other studies have found that humidity and stability effects from BC above and below clouds can cause both increases and decreases in clouds (Koch and Del Genio, 2010). The IPCC did not assign a sign to the net forcing of the semi-direct effect (Denman et al., 2007).

More recently, Koch and Del Genio (2010) found in their review of the literature that most model studies generally indicate a global net negative semi-direct effect (i.e., the effect of atmospheric heating by absorbing aerosols on cloud formation and lifetime causes net cooling). This was observed despite regional variation in the cloud response to absorbing aerosols (such as BC), and resulting regional differences in warming and cooling from the semi-direct effect. In contrast to Koch and Del Genio, Jacobson (2010) found that the semi-direct effect is positive. This difference may be due in part to the more inclusive definition used by Koch and Del Genio. Isaksen et al. (2009) reported a range of -0.25 to +0.50 W m⁻².

While the sign of the semi-direct effect is therefore in question, the glaciation effect is very likely a warming effect, though it occurs only in some mixed-phase clouds. This indirect effect is caused by BC aerosols (and some other particles such

as mineral dust) serving as ice nuclei in a super-cooled liquid water cloud, thereby enabling precipitation rather than delaying it (Denman et al., 2007; Lohmann and Hoose, 2009). However, some preliminary work (e.g., Penner et al., 2009) suggests that the effect of soot on ice nuclei and their subsequent effect on cirrus clouds could offset some of the warming resulting from the glaciation effect.

Most estimates of the forcing from aerosol indirect effects are based on all aerosol species (e.g., total PM) and are not estimated for individual species (e.g., BC alone). The net indirect effect of all aerosols is estimated as a negative value. The IPCC (Forster et al., 2007) estimated the change in cloud albedo due to all aerosols to have a radiative forcing of -0.7 W m^{-2} , with a 5 to 95% confidence range of -0.3 to -1.8 W m^{-2} and a low level of scientific understanding.¹¹ A recent study by Bauer and Menon (2012) found a smaller net cloud interaction effect due to aerosols of only -0.1 W m^{-2} .¹²

There are some BC-specific effects that do not fall into the IPCC categories from Table 2-4, which have been identified in a limited number of studies. The first is called the "cloud absorption effect." This effect is the result of BC particles being included between and within cloud particles, increasing absorption (or decreasing reflectivity) of the clouds. Jacobson (2010) suggests that the inclusion of this cloud absorption effect may increase warming from BC by as much as 75%. Additionally, recent experimental work above the Amazon indicates that cloud albedo will increase with increasing aerosol loading (as predicted by the cloud albedo effect) only as long as the aerosol loading is smaller than a critical threshold. Above this threshold, additional aerosols from biomass burning actually lead to a decrease in cloud albedo, attributed to inhibition of cloud formation by absorbing aerosols (Ten Hoeve et al., 2011; Koren et al., 2008). The second effect that does not fall into the more common IPCC categories is called the "BC-water vapor effect" (Jacobson, 2006, 2010). This effect is based on an increase in water vapor due to inhibition of precipitation from clouds, related to the cloud absorption effect. However, BC can also lead to a decrease in relative humidity in some regions where the increase in atmospheric temperature at altitude does not lead to increased evaporation at

the surface. It is unclear what the magnitude of this effect is.

To summarize, it is unclear to what extent BC contributes to the overall aerosol indirect effect. As a result, this report does not assign any central estimate or even a range of possible values for the role of BC in the overall indirect aerosol effect. BC's role in the first and second indirect effects (cloud albedo and cloud lifetime effects) is likely to be cooling, but possibly to a lesser extent than for other aerosols. Although freshly emitted, externally mixed BC particles are hydrophobic and would be less active cloud condensation nuclei (CCN), aging may increase their ability to serve as CCN (Dusek et al., 2006). Recent work (e.g., Bauer et al., 2010) using models with a more explicit representation of aerosol mixing than older models suggests that the role of BC in the indirect effect may be greater than previously thought. Similarly, BC may also participate in the thermodynamic indirect effect for mixed-phase clouds, but whether this effect is net warming or cooling is still uncertain. BC has a primary role in the semi-direct effect, but this effect may produce warming or cooling depending on conditions. Finally, BC particles may contribute to warming from the glaciation indirect effect in mixed-phase clouds, the cloud absorption effect, and the water vapor effect, but the magnitude of these effects are uncertain. A comprehensive, quantitative estimate of the net effect of BC would require an assessment of the likely bounds of these cloud effects. While this EPA report does not assess those bounds, the UNEP/WMO assessment (UNEP and WMO, 2011a) found that the most likely case was that the cloud interaction effects resulting from BC alone would cancel out (mean value of 0.0 W m^{-2}), but with a central range of -0.4 to $+0.4 \text{ W m}^{-2}$. The forthcoming IGAC/SPARC study will present a more definitive bound on these effect estimates. Studies are just beginning to estimate indirect or semi-direct radiative effects for BC at a regional level (e.g., Bauer and Menon, 2012).

2.6.1.4 Snow and Ice Albedo Forcing

BC deposited on snow and ice leads to positive radiative forcing. It darkens the surface which decreases the surface albedo, and it absorbs sunlight, heating the snow and ice (Warren and Wiscombe, 1980). The snow and ice albedo effect is strongest in the spring because snow cover is at its greatest extent, and spring is a season with increased exposure to sunlight (Flanner et al., 2009). BrC has also been found to contribute to snow and ice albedo forcing (Doherty et al., 2010). Chapter 5 also addresses observations of BC in snow in more depth.

¹¹ The IPCC definition of "level of understanding" is a qualitative measure based on a combination of the quantity of evidence available and the degree of consensus in the literature.

¹² Bauer and Menon estimated the indirect cloud albedo effect as -0.17 W m^{-2} and the semidirect effect as -0.10 W m^{-2} , but noted these effects "can be isolated on a regional scale, and they often have opposing forcing effects, leading to overall small forcing effects on a global scale."

There are a number of estimates of the magnitude of radiative forcing due to the snow albedo effect (see Figure 2-14). In a modeling study, Hansen and Nazarenko (2004) estimated the global average radiative forcing of BC on snow and ice to be +0.16 W m⁻² for what they considered to be the most realistic of the four cases that were simulated in their study. In later work, Hansen et al. (2007b) lowered this estimate to +0.05 W m⁻², with a probable range of 0 to +0.1 W m⁻² (Hansen et al., 2007b). Relying on these studies, the IPCC (Forster et al., 2007) adopted a best estimate for the global average radiative forcing of deposited BC on snow and ice of +0.10 (±0.10) W m⁻², although the authors acknowledged a low level of scientific understanding regarding this effect. In more recent work, Flanner et al. (2007) estimated the average forcing of BC on snow and ice (from fossil fuels and biofuels) at +0.043 W m⁻², of which +0.033 W m⁻² was attributed to BC from fossil fuels. When biomass burning was included in the calculation, the forcing of BC on snow and ice was estimated to be approximately +0.05 W m⁻². Bond et al. (2011) estimated a global forcing of +0.047 W m⁻², of which 20% was calculated to occur in the Arctic (defined as north of 60 degrees), and suggested that more mechanistic studies in general yield estimates lower than the central IPCC estimate of +0.1 W m⁻². In line with these more recent studies, the UNEP/WMO assessment (UNEP and WMO, 2011a) estimated 0.05

W m⁻², with a range of 0.01 to 0.10 W m⁻², for the snow and ice albedo effect.

Hansen et al. (2007b) also investigated the “effectiveness” (or “efficacy”) of the snow albedo forcing. This is a relative measure of positive feedback effects that occur with BC, compared to the feedbacks that occur with warming due to CO₂ forcing. They calculated that the radiative forcing from decreases in surface albedo is 2.7 times more effective at warming than radiative forcing from CO₂. This is a result of the energy absorption from the BC being directly applied to melting snow rather than spread throughout the height of the atmosphere. BC particles left behind in melting surface snow can concentrate and further reduce the surface albedo (see section 5.6). Furthermore, BC deposited on ice and snow will continue to have radiative effects as long as the BC remains exposed (i.e., until the snow melts away or fresh snow falls). Melting snow can expose a dark surface, leading to a positive feedback. Flanner et al. (2007) found a larger efficacy of 3.2, with an uncertainty range of 2.1 to 4.5. Based on the more recent work of Flanner (2009) and Koch (2009), the UNEP/WMO assessment (UNEP and WMO, 2011a) adopted an even larger effective forcing of a factor of five. Flanner et al. (2011) also found that observed Northern Hemisphere snow retreat between 1979 and 2008 (from all causes)

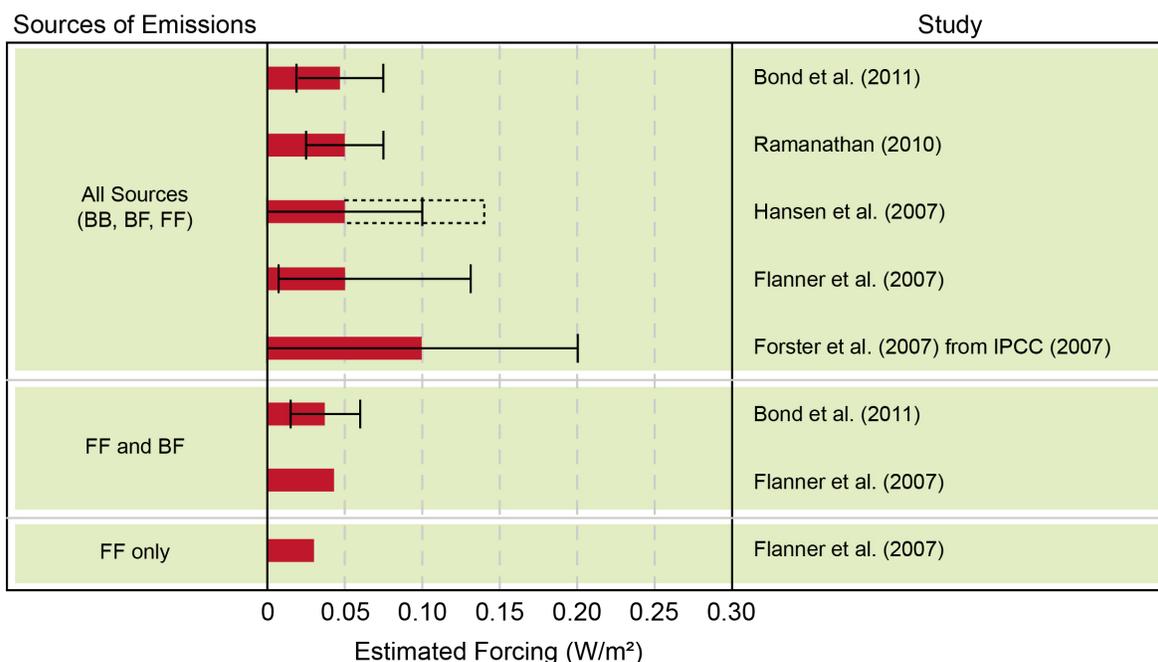


Figure 2-14. Estimates of Snow and Ice Albedo Radiative Forcing Effects from BC Emissions Only. Note: BB = open biomass burning; BF = biofuels; and FF = fossil fuels. Dashed lines indicate the estimated range of snow and ice albedo radiative forcing when forcing efficacy is considered. Hansen et al. (2005) estimate an efficacy of 170% for the snow and ice albedo radiative forcing effect of BC emissions. (Source: U.S. EPA)

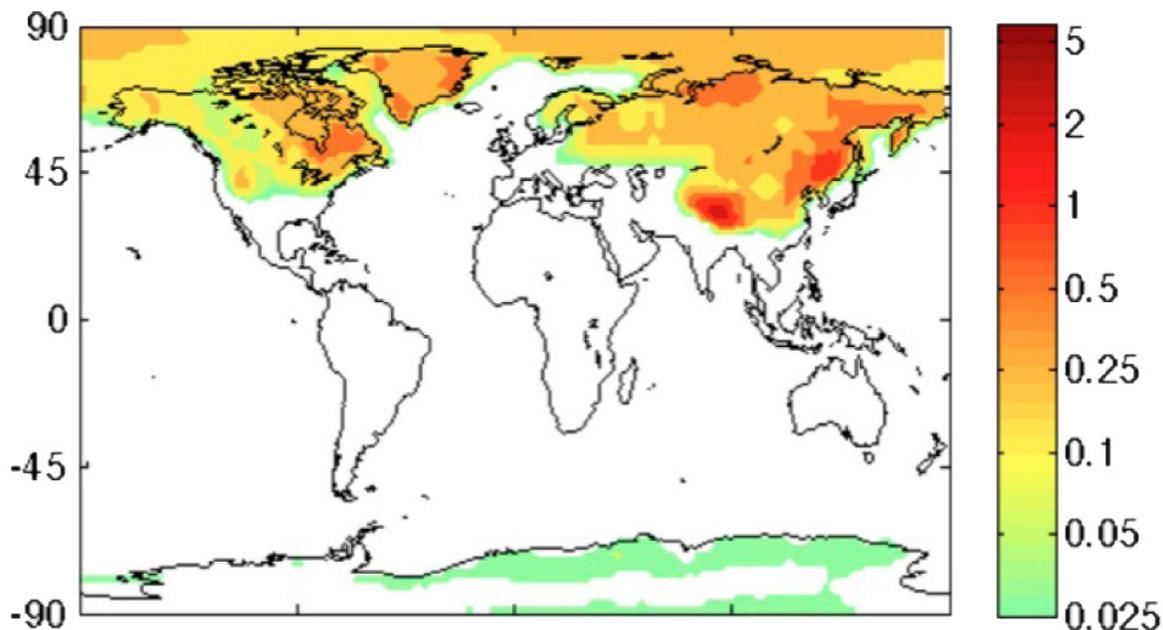


Figure 2-15. Snow and Ice Albedo Forcing by BC, simulated with the Community Atmosphere Model. (Bond et al., 2011)

would be consistent with a total albedo feedback on the order of $+0.45 \text{ W m}^{-2}$. This suggests that albedo feedback is a larger process than represented in most climate models.

For snow and ice, there is evidence that all atmospheric PM, including all mixtures of BC and OC, increases the net solar heating of the atmosphere-snow column (Flanner et al., 2009). This means that mixtures of BC and OC that are transported over snow-covered areas may have a net warming influence regardless of the ratio of the two compounds (although Flanner et al. did not include cloud effects). This is in contrast to direct radiative forcing estimates, which are strongly influenced by the ratio of BC to other cooling PM components such as OC. Flanner et al. (2009) also found that fossil-fuel and biofuel BC and OM emissions contributed almost as much to springtime snow loss in Eurasia as did anthropogenic CO_2 . The size and composition of the deposited particles affects how long they remain on or near the surface where they are able to reduce albedo.

Snow and ice albedo forcing is confined to areas with snow and ice cover (approximately 7.5%-15% of Earth's surface; see section 5.6). Thus, global average forcing estimates do not convey the significant spatial and temporal variability in the radiative forcing of BC on snow and ice. Radiative forcing from changes in snow and ice albedo from BC are estimated to be much larger than the global

averages for much of Northern and Eastern Europe, Russia, and China. These effects are especially pronounced in the Arctic and the Himalayas. Flanner et al. (2007) calculated an average forcing of BC on snow and ice of $+1.5 \text{ W m}^{-2}$ in the Tibetan plateau, with instantaneous forcings¹³ of up to $+20 \text{ W m}^{-2}$ in the spring. These high values are due to the large amount of mountain snow and ice cover as well as the proximity to high emissions of BC from parts of China and the Indian subcontinent. Large radiative forcing values have also been estimated over the Arctic. Hansen and Nazarenko (2004) calculated an average forcing due to BC on snow and ice of $+1 \text{ W m}^{-2}$ in the Arctic compared to $+0.3 \text{ W m}^{-2}$ over the Northern Hemisphere as a whole. However, these estimates are based on global numbers that were reduced by a factor of three in later papers (Hansen et al., 2007b). The full spatial distribution of forcing by BC on snow and ice as simulated by Bond et al. (2011) is shown in Figure 2-15. The effects of BC on the Arctic and the Himalayas are described in more detail in sections 2.6.4 and 2.6.5, below.

2.6.1.5 The Radiative Forcing Effects of OC and other Co-Pollutants

Although BC is mixed with other pollutants, both at the point of emission and in the atmosphere, most studies examine the impact of different types

¹³ Instantaneous radiative forcing refers to the flux at the tropopause, rather than forcing averaged over a longer time period.

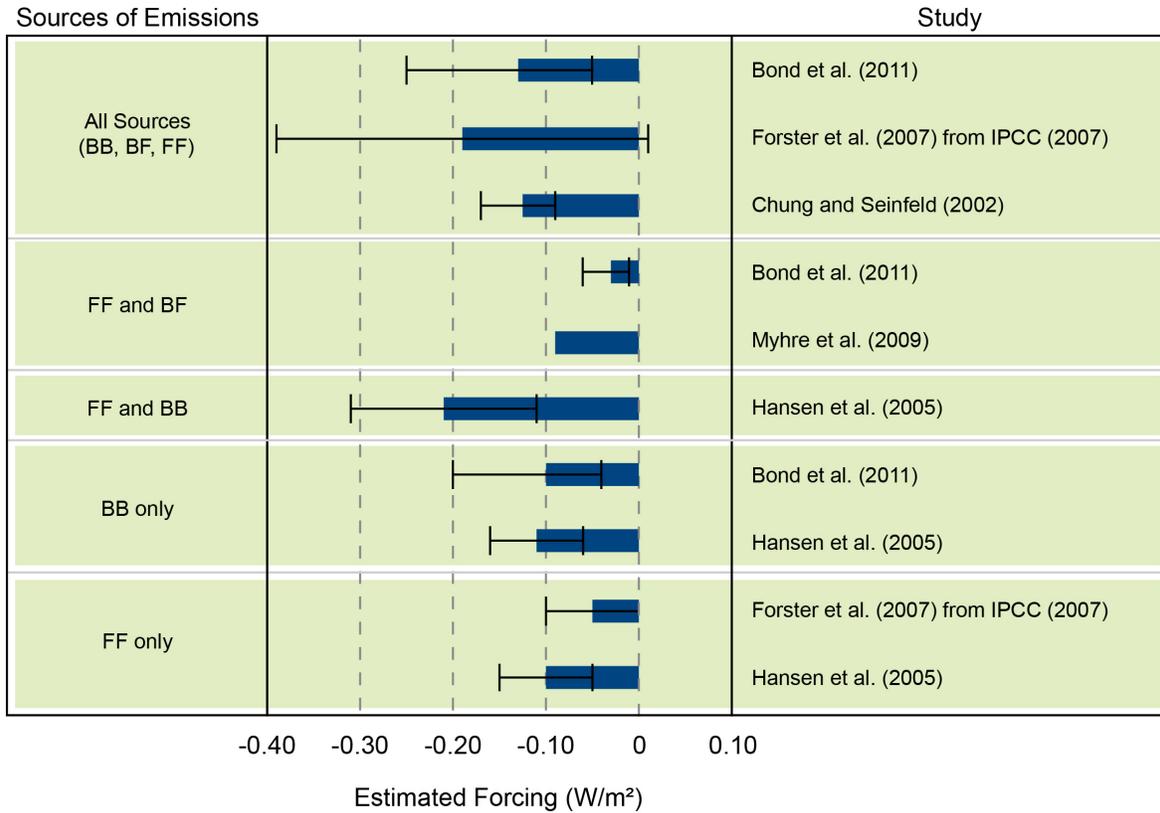


Figure 2-16. Estimates of Direct Radiative Forcing from OC Emissions Only. Note: BB = open biomass burning; BF = biofuels; and FF = fossil fuels. (Source: U.S. EPA)

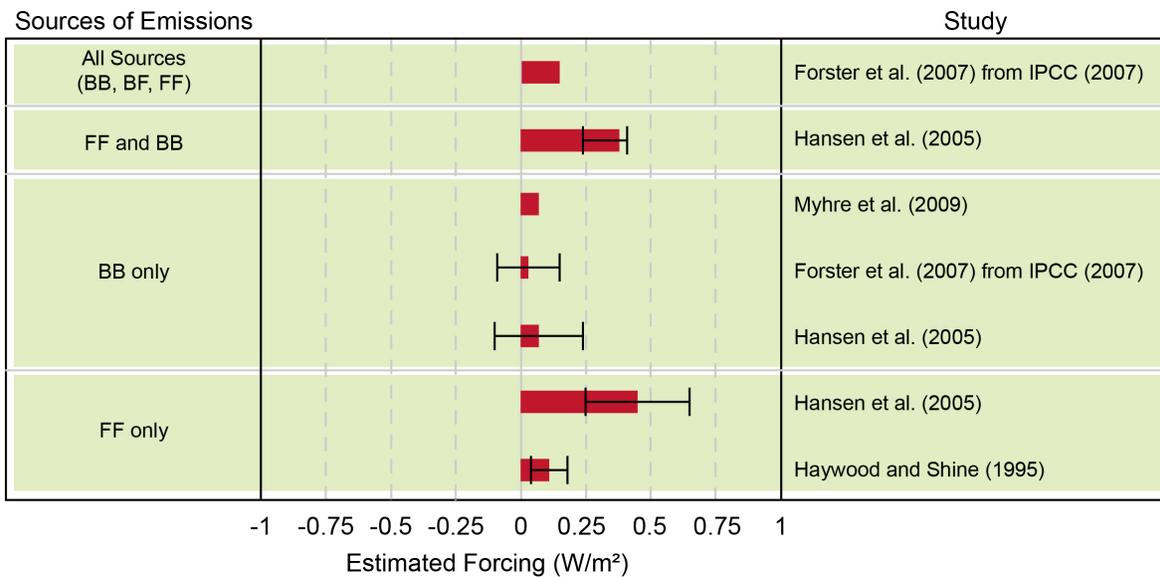


Figure 2-17. Estimates of Direct Radiative Forcing from BC and OC Emissions. Note: BB = open biomass burning; BF = biofuels; and FF = fossil fuels. Forster et al. (2007) – from IPCC (2007) – estimate the uncertainty surrounding estimates of direct radiative forcing from BC and OC independently. For this reason, the uncertainty surrounding the combined estimated direct radiative forcing from BC and OC emissions from all sources according to Forster et al. (2007) is omitted. (Source: U.S. EPA)

of aerosols in isolation. Only a limited number of studies consider the impacts of co-pollutants, and most of these studies have focused on OC rather than all aerosol species. Figure 2-16 shows estimates of direct radiative forcing for OC from a number of studies. As indicated in Figure 2-16, OC emissions from all sources are estimated to have net cooling impacts. For example, the IPCC (2007) estimated the negative direct radiative forcing of OC aerosols from all sources at $-0.19 (\pm 0.20) \text{ W m}^{-2}$ and from fossil fuel alone at $-0.05 (\pm 0.05) \text{ W m}^{-2}$.

When OC and BC emissions are combined, the estimates of global average direct radiative forcing are generally positive. Figure 2-17 shows estimates for BC and OC combined from different sources. Here, the total direct radiative forcing from BC and OC emissions from all sources was estimated by IPCC (2007) at approximately $+0.15 \text{ W m}^{-2}$ (global average), and even biomass burning aerosols were estimated to have a positive net forcing of $+0.03 (\pm 0.12) \text{ W m}^{-2}$. Another study that calculated a net forcing from BC and OC from all sources reported a net global average forcing of about $+0.27 \text{ W m}^{-2}$ (Bond et al., 2011). The UNEP/WMO assessment reports a central net value of BC and OC forcing (including the snow albedo efficacy factor) of 0.41 W m^{-2} (UNEP and WMO, 2011a).

Several additional factors must be taken into consideration in interpreting these estimates. First,

it is important to note that like BC, OC exhibits high spatial variability in direct forcing effects (see Figure 2-18). The regions of highest direct forcing by OC may not coincide with regions of highest direct forcing by BC (see Figures 2-13 and 2-15 for comparison). In addition, most studies evaluating the net effects of BC and OC do not consider indirect effects, and inclusion of these effects will change the net forcing estimates. One study, Chen et al. (2010), found that for one scenario reducing BC and OC in a 2 to 3 ratio, the aerosol indirect effects were larger than (and opposite in sign to) the direct effects. In addition, studies looking at forcing effects due to OC generally consider primary OC emissions only. Secondary organic aerosols (SOA), however, can also make a substantial contribution to the organic aerosols. SOA arises from the oxidation of gaseous VOCs. More recently, Robinson et al. (2007) proposed a more dynamic evolution of aerosol OC in the atmosphere. Based on measurements and models, they suggested that low volatility organic compounds, which are emitted as PM, evaporate, oxidize, and condense over time. The semi-volatile nature of the primary emission of OC may have additional implications for our understanding of OC and OC to BC ratios on climate, although this remains poorly understood (Jimenez et al., 2009).

The inclusion of other species, mainly nitrate and sulfate aerosols, also tends to reduce the estimate of net forcing. In particular, the presence or absence

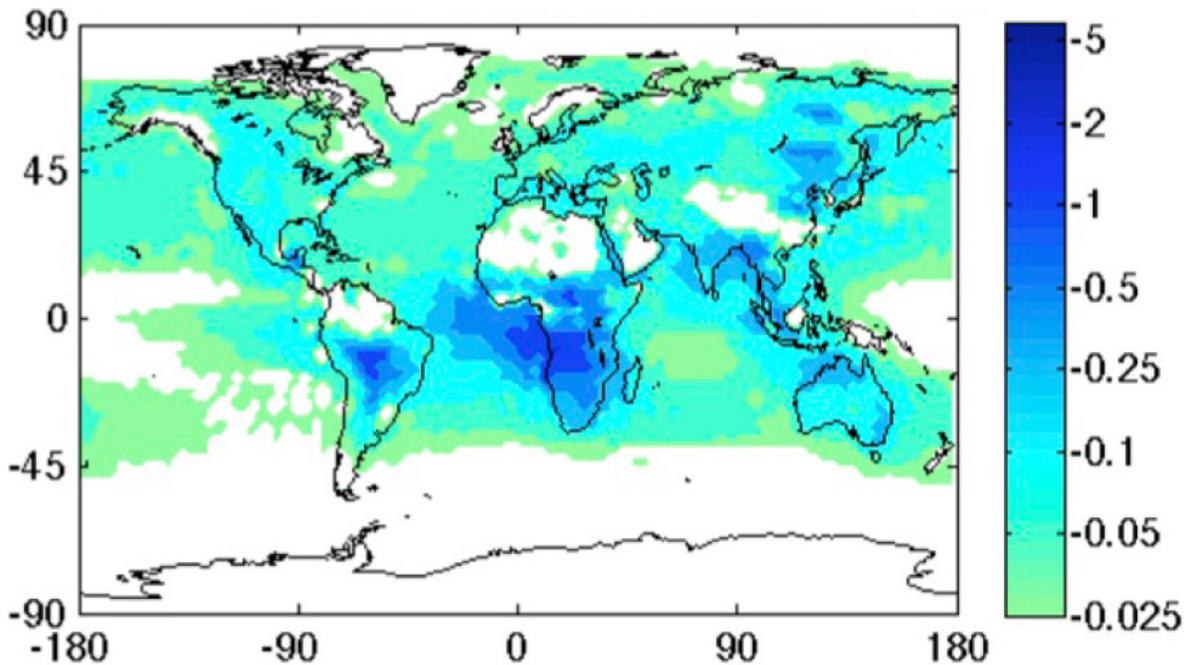


Figure 2-18. Direct Forcing by OC from All Sources, simulated with the Community Atmosphere Model. (Bond et al., 2011)

of sulfates and nitrates, which together comprise a large fraction of aerosol mass, in calculations of indirect effects can dominate radiative forcing calculations. Inclusion of both direct and indirect effects of aerosol species in the review by Ramanathan and Carmichael (2008) led to an estimate of the total aerosol effect including direct and indirect effects of -1.4 W m^{-2} , in contrast to a calculated BC direct forcing of $+0.9 \text{ W m}^{-2}$. However, because much of the nitrate and sulfate precursor emissions come from sectors that are not rich in BC, the net global effect of aerosols can be less important than the estimates of the net effects of aerosols from a specific sector or measure (discussed further in section 2.6.1.6). These aerosols also play a role in the mixing state and therefore the direct radiative forcing effect of BC, as discussed in sections 2.5 and 2.6.1.2. Therefore, ambient concentrations of these other aerosols can be important in determining the influence of BC reductions. Using surface and aircraft measurements, Ramana et al. (2010) found that the ratio of BC to sulfate was important in determining the net warming or cooling impact of pollution plumes in China.

2.6.1.6 Sector-Based Contributions to Radiative Forcing

As described in Chapter 4, BC emissions can be attributed to a wide range of sectors (e.g., transportation, residential, industrial, and biomass burning). Some studies have attempted to quantify the radiative forcing effect of emissions mixtures containing BC and other co-pollutants by estimating the radiative forcing of defined emissions sectors. Comparisons among studies, however, is hindered in part by variation in several parameters, including estimates of the sector-level contributions, the relative

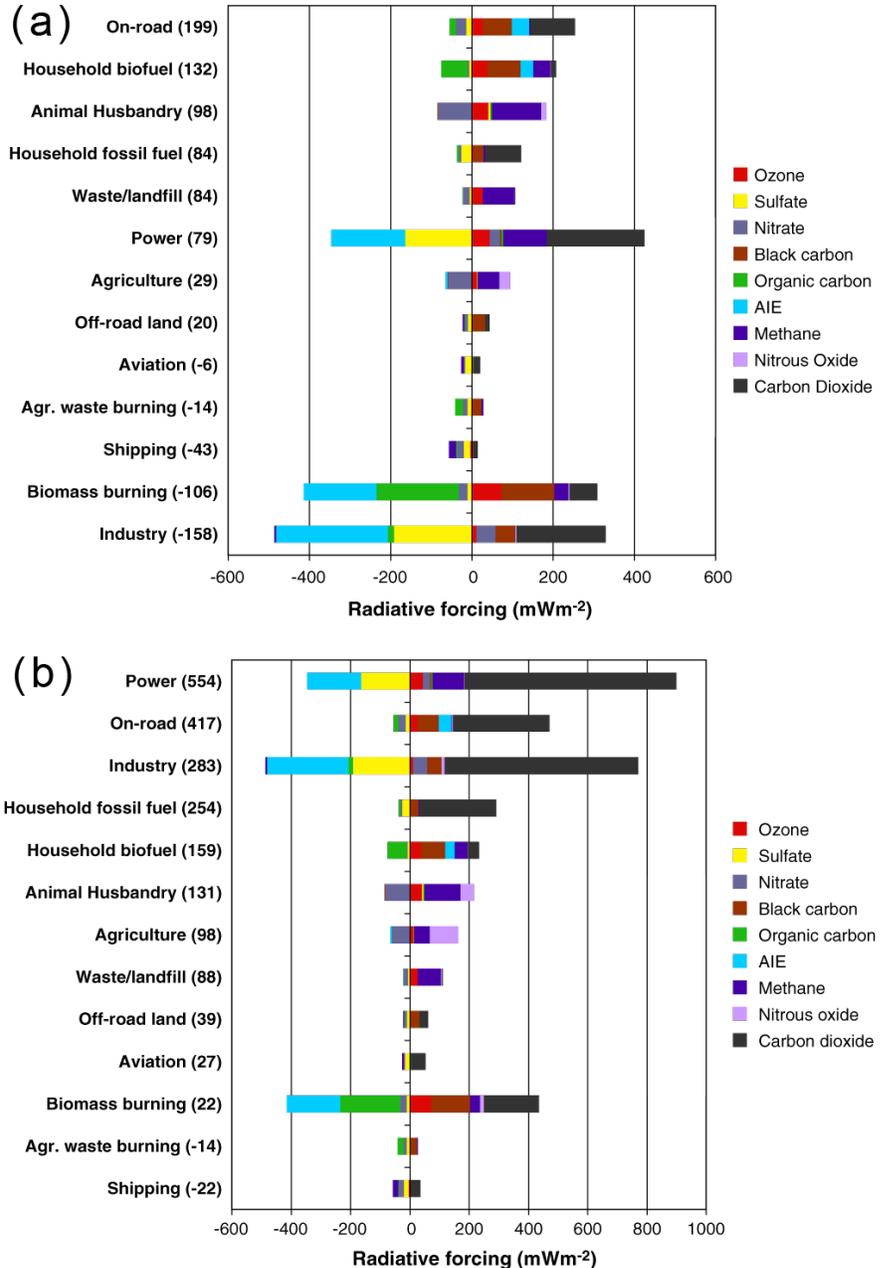


Figure 2-19. Global Radiative Forcing Due to Perpetual Constant Year 2000 Emissions, Grouped By Sector, at (a) 2020 and (b) 2100 and Showing the Contribution from Each Species. The sum is shown on the title of each bar, with a positive radiative forcing means that removal of this emissions source will result in cooling. AIE is the aerosol indirect effect. (Unger et al., 2010)

fraction of warming and cooling aerosols, and the microphysical properties of these aerosols.

Unger et al. (2010) examined the impacts of sector-specific emissions on the short- and long-term radiative forcing from a range of pollutants. Figure 2-19 shows that the mixture of

emissions from some of the largest BC emissions sources contributes considerably to total radiative forcing. On-road transportation emissions are the largest contributor to radiative forcing in the short term (by 2020), due to a combination of GHG and BC emissions. On-road transportation is also seen to be the second largest contributor in the long term (by 2100), but this is largely the result of the significant GHG emissions from this sector. Residential biofuel combustion is the second largest contributor in the short term due to the contribution from BC and CH₄. Since these sources have lower net GHG emissions, they contribute less to total global radiative forcing in the long term. However, these calculations have substantial uncertainties owing to the details of aerosol physics and chemistry, the interactions of aerosols and clouds, and the regional nature of the radiative forcing, as discussed earlier in this chapter. Bauer and Menon (2012) focused on regional differences in the impact of emissions from different source categories, and concluded that the largest opportunities to reduce positive forcing due to all aerosols included transportation in all regions, agricultural burning in Europe and Asia, and residential cooking and heating (“domestic sector”) in Asia.

There is significant disagreement regarding the net impact of aerosol emissions from open biomass burning on radiative forcing. As noted in the previous section, the IPCC estimated the net direct radiative forcing impact from open biomass burning aerosols to be small, but positive at +0.03 (±0.12) W m⁻² (Forster et al., 2007). However, because of uncertainties regarding the extent and composition of emissions from this source category, and the indirect radiative forcing effects of biomass burning aerosols, it is not clear if this sector has an overall global warming or cooling effect. Kopp and Mauzerall (2010) developed probability distributions from multiple studies to examine the likelihood of warming from individual sectors. Based on existing evidence, they concluded that open burning in forests and savannas is unlikely to contribute to warming, while the effect of open burning of crop residues remains uncertain. The results of current analyses are sufficiently different that there is no consensus on the likelihood of warming. Stohl et al. (2007) concluded that biomass burning has a “significant impact on air quality over vast regions and on radiative properties of the atmosphere” and in particular “has been underestimated as a source of aerosol and air pollution for the Arctic, relative to emissions from fossil fuel combustion.” As discussed further in section 5.6, surface snow records indicate that biomass burning is currently a major source of BC in Greenland and the North Pole (Hegg et al., 2010). Bauer and Menon (2012) point to agricultural

burning in Europe and Asia as contributing to positive net forcing in those regions. Additional work is needed to improve scientific understanding of the radiative forcing impacts of open biomass burning.

Several modeling experiments, such as Jacobson (2002, 2005, 2010), Hansen et al. (2005), and Schulz et al. (2006), and observationally constrained studies such as Ramanathan and Carmichael (2008), have found that carbonaceous aerosols from biofuel combustion and fossil fuel combustion both contribute to warming. Among fossil fuels, diesel combustion for transportation is the largest contributor to global BC emissions and several studies suggest these emissions may contribute to warming (Jacobsen 2002, 2005; Hansen et al., 2005). Kopp and Mauzerall (2010) concluded that carbonaceous PM emissions from gasoline combustion are unlikely to contribute to warming,¹⁴ while diesel combustion and residential coal combustion are very likely to contribute to warming.

Kopp and Mauzerall (2010) also found mixed results with respect to the contribution of residential biofuel combustion for the models included in their assessment. More recent work by Rehman et al. (2011) has reported higher BC concentrations due to burning biomass for cooking than previously reported, both indoors and outdoors, in the study region in northern India. Moreover, Rehman et al. found that the albedo of the particles in the study villages indicated high absorption. Wavelength analysis suggested that though OC concentrations were a factor of five higher than BC concentrations, the OC included significant absorbing BrC.

Aviation is also a source of BC emissions. While the amount of BC emitted by aircraft at cruise altitudes is subject to large uncertainties (see Appendix 2-10), BC particles at these altitudes absorb not only the downward radiation but also the reflected upward radiation. In addition, intercontinental flight tracks are concentrated in the arctic stratosphere and particles emitted in this region may be deposited in the Arctic ice and snow. Research is needed to quantify radiative forcing due to BC of aviation origin.

A few studies highlight the substantial uncertainties regarding the contribution of biofuel combustion

¹⁴ For gasoline vehicles, it should be noted that the introduction of new engine technologies (e.g., some types of gasoline direct injection) in recent model years has increased BC/PM ratios in some new gasoline-powered motor vehicles (Smallwood et al., 2001), which may change the warming profile of emissions from these vehicles. See Chapter 8 for more discussion of this issue.

and fossil fuel combustion to warming, given our limited understanding of how carbonaceous aerosols affect cloud processes. In the modeling experiments by Chen et al. (2010), reductions in fossil fuel carbonaceous aerosols (BC and OC) lead to decreases in CCN, leading to a decrease in cloud albedo, causing an increase in radiative forcing. The impact of these cloud changes equal or exceed the direct radiative forcing impacts. This result contrasts with that of Jacobson (2010) and Bauer et al. (2010) in which estimated warming from indirect effects did not exceed the direct and other radiative forcing from fossil fuel emissions. The distinction between biofuel combustion and fossil fuel combustion, in terms of the effect on radiative forcing, is particularly noticeable in the residential sector. Anan et al. (2009) estimate that the global annual mean radiative forcing from BC from residential fuel consumption in Asia is positive for both biofuel and fossil fuel consumption, but that the net effects on radiative forcing from residential biofuel consumption in Asia (accounting for BC and the range of other co-emitted pollutants) is negative.

Fossil fuels burned for electricity generation contribute only a small fraction of carbonaceous aerosol emissions, though this sector is a large source of long-lived, warming GHGs and short-lived cooling sulfate aerosols (Shindell and Faluvegi, 2009). Thus, though their study found that the sector is the largest single contributor to warming on the 100-year time scale, this is attributable to GHG emissions rather than emissions of BC.

2.6.2 Impact of BC Radiative Forcing on Temperature and Melting of Ice and Snow

As mentioned in section 2.6.1, global average radiative forcing is linearly related to the global mean temperature at the surface (Forster et al., 2007). Radiative forcing from agents such as BC has similar effects on global mean temperature as radiative forcing from CO₂ and other GHGs (Hegerl et al., 2007), though the efficacy of the forcing may differ slightly (especially for the snow/ice albedo effect). Temperature itself has already been linked to a range of climate impacts as identified in, for example, the 2009 USGCRP report, "Global Climate Change Impacts in the United States." This and other recent climate change assessments describe the risks and impacts associated with climate change, including degradation of air quality, temperature increases, changes in extreme weather events, effects on food production and forestry, effects on water resources, sea level rise, disruption to energy consumption and production, and potential harm to ecosystems and wildlife. Though few studies explicitly link BC to all of these outcomes,

to the extent that BC increases temperature it will contribute to these impacts, especially impairment of air quality and sea level rise (via melting of ice, snow, and glaciers).

Work by Jacobson often uses temperature change as an endpoint, rather than radiative forcing. As discussed earlier in section 2.6, there are advantages and disadvantages to using temperature change as an endpoint. Jacobson also includes a more complete suite of BC and co-pollutant effects than most other models. Jacobson (2010) found that the net effect of existing fossil fuel BC plus OC emissions was to warm the climate by 0.3 to 0.5°C at equilibrium compared to a case without those emissions.

There have been some efforts to translate regional direct radiative forcing estimates into regional changes in temperature. For example, Chung and Seinfeld (2005) used the GISS GCM model with BC emissions from Bond et al. (2004) to predict annually averaged changes in regional temperatures due to BC direct radiative forcing, based on simulations of 100 years of future emissions and temperature responses. They predict that externally mixed BC leads to an increase in average annual surface temperatures of 0.29°C in the Northern Hemisphere and 0.11°C in the Southern Hemisphere, when model results for the latter 75 years of simulations are averaged (the first 25 years of simulations were omitted from the results, to allow for the model to settle). Internally mixed BC is predicted to result in an increase in average annual surface temperatures of 0.54°C in the Northern Hemisphere and 0.20°C in the Southern Hemisphere over the same period. Despite the uncertainties regarding mixing state, the authors show that surface temperature response to regional direct radiative forcing is more concentrated in the Northern Hemisphere, especially in locations of high latitude.

Few studies have evaluated the North America-specific temperature impacts associated with BC emissions. However, Qian et al. (2009) found that BC emissions lead to warming of a tenth to a full degree Celsius over snow in the western United States. Simulations show that BC absorption of solar radiation in the atmosphere leads to as much as 0.6°C of warming in the lower and mid troposphere over most of North America, including the Arctic region (Ramanathan and Carmichael, 2008). Additional estimates of regional temperature effects associated with BC emissions in the Arctic and the Himalayas are discussed in sections 2.6.4 and 2.6.5, respectively.

Chapter 2

The snow/ice albedo effects of BC have been linked to accelerated melting of snow and ice. While many glaciers around the world and Arctic sea ice have receded in recent decades, attribution of this melt to BC is challenging due to other global and local contributions to warming and precipitation changes. Regardless of the deposited BC, other factors – including the solar zenith angle, cloud cover, snow grain size, and depth of the snow – also influence the snow and ice albedo (Wiscombe and Warren, 1980). The most common method of determining the contribution of BC to accelerated snow and ice melt has been to compare model runs with and without BC influences, and evaluate with observations. Direct measurements are generated by melting and then filtering samples of snow and ice. The filters provide an estimate of BC concentration by comparing their observed optical transmissions to optical transmissions of known amounts of BC (Noone and Clarke, 1998; Warren and Clarke, 1990). The mass is then used to estimate or compare to measured snow albedo, calculating the influence of BC. Another approach has been to apply a known amount of soot to an area, and then compare the measured albedo and melting rate to a nearby clean plot of snow.

In the western United States, deposition of BC on mountain glaciers and snow packs produces a

positive snow and ice albedo effect, contributing to the melting of snowpack earlier in the spring and reducing the amount of snowmelt that normally would occur later in the spring and summer (Hadley et al., 2010; Koch and Del Genio, 2010). This has implications for freshwater resources in regions of the United States that are dependent on snow-fed or glacier-fed water systems. In the Sierra Nevada mountain range, Hadley et al. (2010) found BC at different depths in the snowpack, deposited over the winter months by snowfall. In the spring, the continuous uncovering of the BC contributed to the early melt. A model capturing the effects of soot on snow in the western United States shows significant decreases in snowpack between December and May (Figure 2-20, Qian et al., 2009). Snow water equivalent (the amount of water that would be produced by melting all the snow) is reduced by 2-50 millimeters (mm) in mountainous areas, particularly over the Central Rockies, Sierra Nevadas, and western Canada. In addition, dust deposition on snow, at high concentrations, can have similar effects to BC. A study done by Painter et al. (2007) in the San Juan Mountains in Colorado observed a decrease in snow cover duration of 18-35 days as a result of dust transported from non-local desert sources. As the authors note, in the future, exacerbated dryness in desert and arid regions—

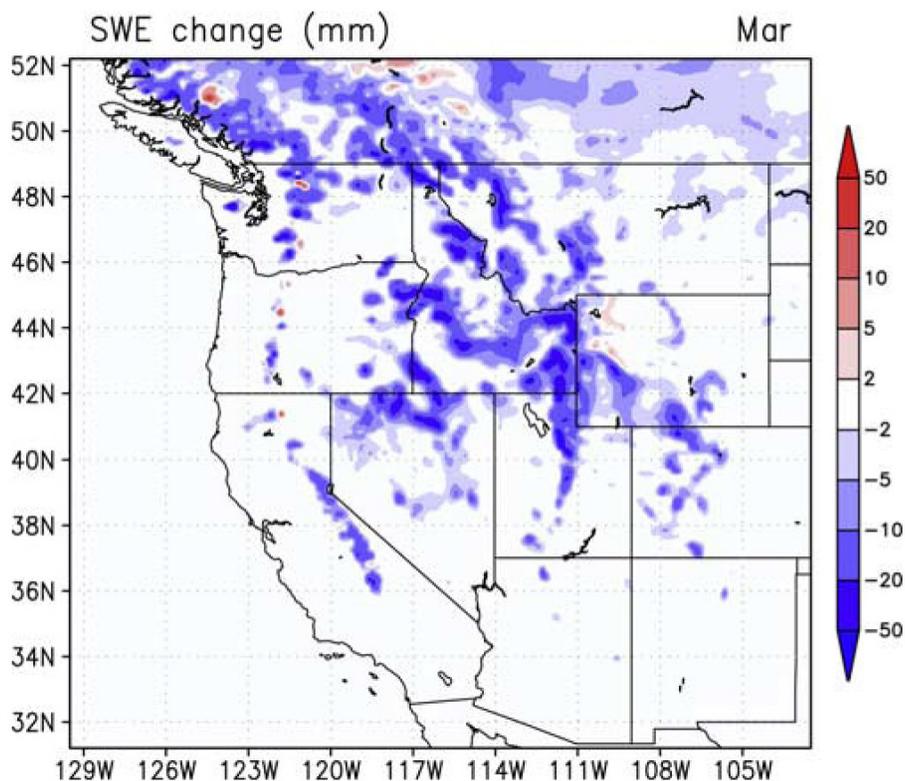


Figure 2-20. Spatial Distribution of Change in Mean Snow Water Equivalent (SWE, mm) for March. (Qian et al., 2009)

especially in the Southwest United States, the Middle East, and the Sahel—could cause increased deposition of dust on mountain snow cover in areas such the Rocky Mountains, the HKHT, and the Alps, respectively, which could lead to considerable reductions in snow cover duration over large areas.

Changes in snow and ice melt due to BC can affect various types of surfaces and geographic locations throughout the world, including Arctic ice caps and sea ice, glaciers, and mountain snowpack (see section 2.6.4 for more detailed treatment of Arctic impacts). For example, Ming et al. (2009) suggest that reduced albedos in some glaciers in west China from BC deposition might accelerate melting of these glaciers. Figure 2-21 shows a Chinese glacier and the concentration of BC that results from melting the upper layers of the snowpack until it is buried by fresh snowfall.

It is important to note that the impacts of BC on snow and ice albedo are not constrained to regions of high elevation or high latitude. BC deposition can also contribute to accelerated melting of seasonal, non-mountain snow, especially in mid-latitude regions due to the additional exposure to sunlight these regions experience (compared to polar regions). In addition, these regions are generally closer to BC emissions sources than polar regions are, so concentrations of BC on seasonal snow

accumulations can be considerable (Huang et al., 2011).

2.6.3 Other Impacts of BC

In addition to warming and cooling effects due to absorption and reflection of light both directly and through cloud interactions, BC and other aerosols contribute to climate change through surface dimming and changes in precipitation patterns. The following sections provide information on BC’s contributions to these two types of impacts. It is important to note that unlike BC and other aerosols, GHGs are not associated with surface dimming, nor are they linked directly to changes in precipitation. Changes in precipitation from GHGs are mediated through changes in temperature.

2.6.3.1 Surface Dimming Effects

The absorption of incoming solar radiation by BC reduces the amount of solar radiation reaching the Earth’s surface, an effect referred to as surface dimming in many studies (e.g., Forster et al., 2007). This results in cooling at the surface (even though net forcing measured at the TOA may be positive). A number of studies report evidence of global dimming between the 1960s and the 1980s, followed by an increase in the amount of sunlight reaching the surface during the 1990s to

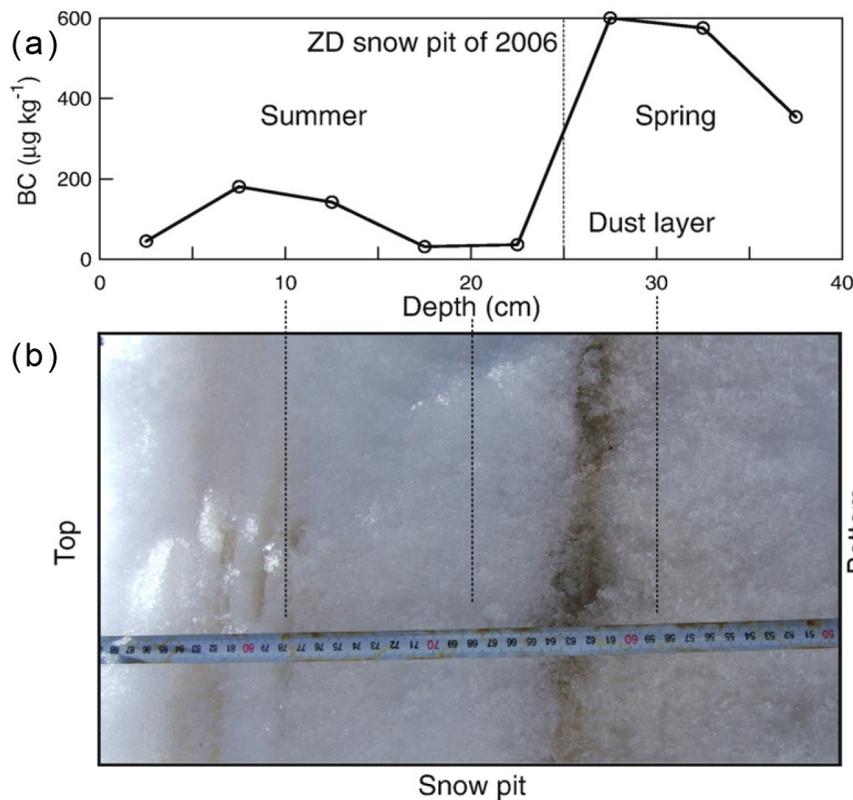


Figure 2-21. BC Concentrations in the ZD Glacier. (a) Measured BC concentration (b) the snow pit of the ZD Glacier. Dust layer at 30 cm indicated the spring (melting season) of 2006. (From Ming et al., 2009, Figure 4)

the present (sometimes referred to as *brightening*) (e.g., see review in Wild, 2009). Numerous studies suggest that the observed dimming and brightening trends are caused by changes in aerosol emissions over time and the interaction of aerosol direct and indirect radiative forcing (Stanhill and Cohen, 2001; Wild et al., 2005; Streets et al., 2006; Ruckstuhl et al., 2008). See Table 2-5 for a summary of how aerosol interactions affect surface dimming.

Estimating the magnitude of the surface dimming effect is complicated, largely because the vertical distribution of BC in the atmospheric column affects the impacts at the surface. Isolating the effect of BC is also difficult due to the interactions of non-BC aerosols. For example, many non-BC aerosols (primarily sulfates) scatter incoming solar radiation, reducing the energy reaching the surface (Dwyer et al., 2010; Ramanathan and Carmichael, 2008). In addition, the indirect effect of aerosols on cloud albedo and cloud lifetime may decrease solar radiation at the surface (Ramanathan and Carmichael, 2008). Furthermore, surface cooling combined with atmospheric heating from BC may increase the stability of the boundary layer (e.g., the bottom layer of the troposphere that is in contact with the surface of the earth) and reduce vertical mixing. This increase in atmospheric stability reduces natural removal processes for air pollutants, resulting in worse air pollution episodes (Ramanathan and Carmichael, 2008).

In some regions, BC, BrC, sulfates, organics, dust and other components combine to form pollution clouds known as ABCs, which have been linked to global dimming (Ramanathan and Carmichael, 2008; Ramanathan et al., 2007). Ramanathan and Carmichael (2008) estimate the total global dimming effect from ABCs to be -4.4 W m^{-2} , with about -3.4 W m^{-2} from the direct effect of aerosols (roughly half of which is attributed to BC) and the remaining

-1 W m^{-2} from the indirect effect. In the Ramanathan and Carmichael study (2008), the -1.7 W m^{-2} of surface dimming from BC was found to be offset by $+2.6 \text{ W m}^{-2}$ of heating in the atmosphere. This resulted in a net TOA forcing estimate from this study of $+0.9 \text{ W m}^{-2}$, as cited in section 2.6.1.2.

The dimming effects due to BC and the other aerosols are not spatially uniform (see Figure 2-22). A number of studies have found that dimming effects are particularly acute in certain regions associated with high aerosol pollution levels and the presence of ABCs. These include certain major urban areas (Ramanathan and Feng, 2009; Trenberth et al., 2007), primarily in South Asia (Ramanathan and Feng, 2009; Ramanathan et al., 2005; 2007). The ABCs which cover large areas in the North Indian Ocean and South Asia can reduce energy at the surface by 5-10% (Ramanathan et al., 2007; Ramanathan and Carmichael, 2008). Some studies have estimated that the dimming associated with ABCs can mask approximately half of the warming that would occur at the surface in the absence of ABCs due to GHGs alone, especially over Asia (Ramanathan et al., 2007; UNEP, 2008a). Another study estimates that surface dimming causes a reduction of approximately 6% in solar radiation at the surface over China and India when compared to pre-industrial values (UNEP, 2008a). The U.S. Global Change Research Program (CCSP, 2009) estimated surface forcing values as low as -10 W m^{-2} over China, India, and sub-Saharan Africa due to elevated optical depth from aerosol emissions in that region.

2.6.3.2 Precipitation Effects

Aerosols affect the processes of cloud and rain droplet formation. Some studies have linked aerosols to reductions in rainfall, but these interactions are not well understood. A summary of the aerosol interactions with clouds that cause

Table 2-5. Overview of the Different Aerosol Indirect Effects and Their Implications for Global Dimming and Precipitation. This table applies to all aerosols, not just BC. Scientific uncertainty is “very low” for all effects except the cloud albedo effect (for which uncertainty is “low”). For descriptions of the effects, see section 2.6.1.3. (Adapted from Denman et al., 2007, Table 7.10b.)

Effects	Sign of Change in Surface Dimming	Potential Magnitude	Sign of Change in Precipitation	Potential Magnitude
Cloud albedo effect	Positive	Medium	NA	NA
Cloud lifetime effect	Positive	Medium	Negative	Small
Semi-direct effect	Positive	Large	Negative	Large
Glaciation indirect effect	Negative	Medium	Positive	Medium
Thermodynamic effect	Positive or negative	Medium	Positive or negative	Medium

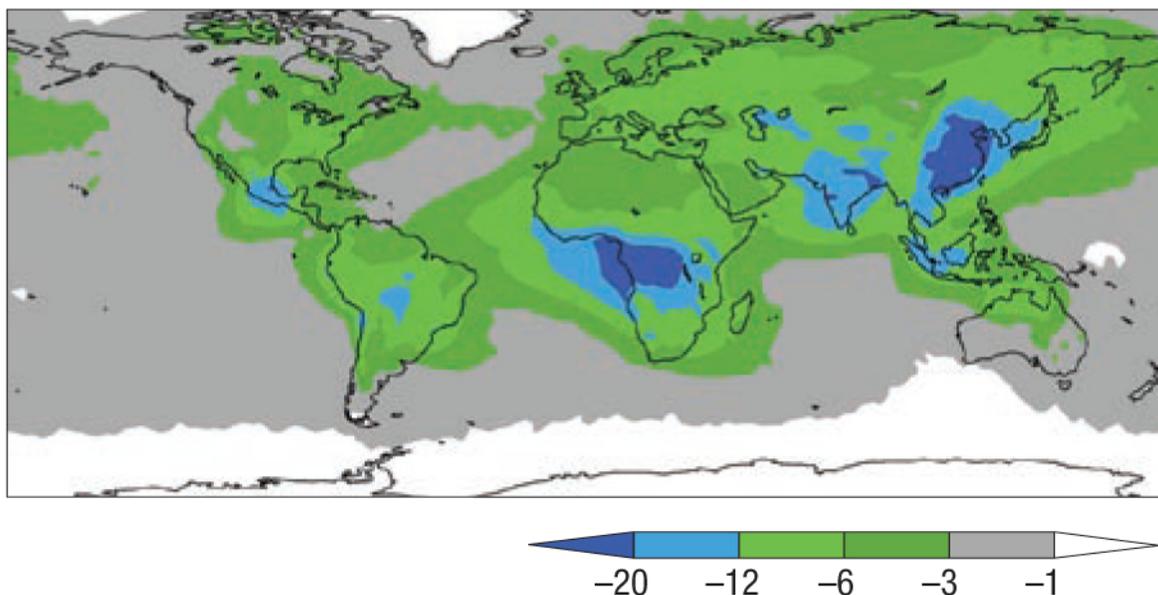


Figure 2-22. Surface Dimming by Anthropogenic Aerosols ($W m^2$). (Adapted from Chung et al., 2005, and Ramanathan and Carmichael, 2008)

changes in precipitation is provided in Table 2-5. According to the IPCC (2007), the precipitation effects on a global scale attributed primarily to BC are from the semi-direct effect (described in section 2.6.1.3) and the increased atmospheric stability resulting from that effect (Ramanathan et al., 2005; Chung and Zhang, 2004; Menon et al., 2002). The increased stability inhibits convection, affecting both rainfall and atmospheric circulation. As discussed in section 2.6.1.3, increased availability of CCN increases cloud lifetime, thereby inhibiting rainfall for a time period, which may be more important for shifting the location of rainfall than changing net global precipitation. There may also be increases in precipitation: BC in particular can stimulate precipitation from ice clouds. However, because of the dependence of precipitation on complex and localized conditions, scientific understanding of these effects is low and models often disagree on the magnitude or sometimes even the sign of changes in precipitation due to factors such as warming or aerosol emissions.

Surface dimming due to all types of aerosols may reduce precipitation by reducing the energy available for evaporation from the Earth's surface (Liepert et al., 2004; Ramanathan et al., 2001). Bauer and Menon (2012) estimate that the global mean change in precipitation due to all aerosols is -0.3%. Because rain is a major removal mechanism for BC from the atmosphere, large decreases in rainfall could result in higher atmospheric concentrations of

BC and other aerosols (Ramanathan and Carmichael, 2008; Ramanathan et al., 2005).

Ramanathan and Feng (2009) suggest that, on a global average basis, reduced precipitation caused by the surface dimming effects of aerosols is likely to be countered with increased precipitation from GHG-induced warming. The effect of aerosols on precipitation, however, varies by area, surface cover, and location. For example, in the tropics, the net effect of aerosols and GHG-induced warming may be reduced rainfall (Ramanathan and Feng, 2009). These shifts in rainfall patterns may have important implications for water availability.

In the United States, Qian et al. (2009) found only small changes in the amount of precipitation in the western U.S. as a result of BC effects. While there is no evidence in North America that links BC or any other specific constituent of PM to changes in precipitation, there are studies that show correlations between total PM emissions and regional precipitation patterns. For example, Bell et al. (2008) find weekly patterns of emissions that correlate with weekly patterns in rainfall in the southeastern United States (Bell et al., 2008). Similar results have also been found for the East Coast of the United States (Cerveny and Balling, 1998).

There is stronger evidence linking aerosols to reduced precipitation in the tropics. Studies have indicated that surface dimming in this region reduces evaporation (Feingold et al., 2005; Yu

et al., 2002a; Hansen et al., 2007b). Other studies have found that the effect in the tropics may be unevenly distributed with increased precipitation just north of the equator (between 0° and 20° N) and decreased precipitation just south of the equator (between 0° and 20° S). This would shift the Intertropical Convergence Zone northward (Chung and Seinfeld, 2005; Roberts and Jones, 2004; Wang, 2004). This northward shift may be caused by the enhanced temperature difference between the Northern and Southern Hemispheres (also a result of global warming from GHGs), which induces a change in circulation and convection in the tropics. Aerosols have also been linked to impacts on regional precipitation in the Amazon basin (Martins et al., 2009; Bevan et al., 2009). This is a region of high biomass burning emissions in the dry season. Further, seasonal biomass emissions have been linked to larger changes in atmospheric circulation patterns by affecting the global distribution of high-level clouds and convection precipitation (Jeong and Wang, 2010). Jeong and Wang (2010) also found that the climate response extends outside of the biomass burning season. The effects of BC aerosols on precipitation may also extend beyond areas of high concentrations. Wang (2007) found the largest change in precipitation occurs in the tropical Pacific region which is far from the regions of largest BC forcing. The effect may be very similar to the pattern of precipitation anomalies associated with the El Niño/Southern Oscillation.

There is also evidence that BC and ABCs slow down the monsoon circulation over South Asia. Specifically, the surface dimming caused by BC aerosols (Meehl et al., 2008) and ABCs (Lau and Kim, 2006; Ramanathan et al., 2005) alters both the north-south gradients in sea surface temperatures and the land-ocean contrast in surface temperatures. These studies estimate an increase in pre-monsoon rainfall during spring followed by a decrease in summer monsoon rainfall, in agreement with observed trends.

Model studies of China have found that BC contributes to increased rainfall in the south and reduced rainfall in the north (Wu et al., 2008; Menon et al., 2002). Wu et al. (2008) simulated the regional climate impacts of BC's direct radiative forcing effect in Asia and found about a 0.6% increase in atmospheric water vapor over southern China, resulting in a precipitation increase of 0.4–0.6 mm/day. In northern China, this study found about a 0.3% decrease in water vapor and a resultant decrease in precipitation. Meehl et al. (2008) found small precipitation increases over the Tibetan Plateau due to BC, but concluded that precipitation over China generally decreases due to BC effects.

2.6.4 BC Impacts in the Arctic

BC emissions that are transported to the Arctic are strongly linked to local warming (Reddy and Boucher, 2007), even if the globally averaged net climate impact of the total particulate emissions from individual sources is uncertain. For example, Quinn et al. (2008) calculated that the contribution of short-lived climate forcers (i.e., CH₄, tropospheric ozone, and tropospheric aerosols, including BC) to Arctic warming is about 80% that of CO₂. BC can have significant snow albedo effects and the magnitude of the cooling effect over snow from co-emitted aerosols is reduced in the Arctic. Because temperature in the Arctic has warmed at twice the global rate over the past 100 years (IPCC, 2007) and because of the dramatic retreat of summer sea ice extent during the satellite observation period (see Figure 2-23), there is interest in mitigation strategies that may slow the near-term rate of climate change in this region.

The estimated radiative forcing from BC is larger over the Arctic than it is on average globally. Due to the lack of sunlight in winter months, the long days in summer, and the increased efficiency of transport of BC emissions from lower latitudes in spring, there is also much larger seasonal variability in the estimates of radiative forcing from BC and other aerosols than there is from GHGs (Quinn et al., 2008). Looking at forcing from fossil fuel and biofuel BC emissions, Quinn et al. (2008) calculated a radiative forcing in the Arctic of +1.2 W m⁻² in the spring, +0.66 W m⁻² in the summer, +0.16 W m⁻² in the fall, and only 0.09 W m⁻² in the winter. Snow albedo forcing in the Arctic was calculated to add an additional +0.53 W m⁻² in the spring, +0.21 W m⁻² in the summer and negligible forcing in autumn and winter. This effect is amplified (e.g., increase in efficacy) by the hastening of the spring thaw that reveals darker ground and water (ocean/lake) surfaces.

Due to the frequency of strong temperature inversions that inhibit atmospheric mixing, and the prevalence of dry conditions that impede wet deposition, the lifetime of aerosol particles in the Arctic is longer than other regions – sometimes weeks, rather than days (see Garrett et al., 2004; Curry, 1995). This leads to a phenomenon known as Arctic haze which is the result of an accumulation of BC, OC, and sulfate particles in the atmosphere above the Arctic (Quinn et al., 2007). Strong surface-based temperature inversions and the dryness of the Arctic troposphere inhibit removal of particles via deposition. Over a highly reflective surface like the Arctic, BC particles absorb solar radiation and warm the atmosphere above

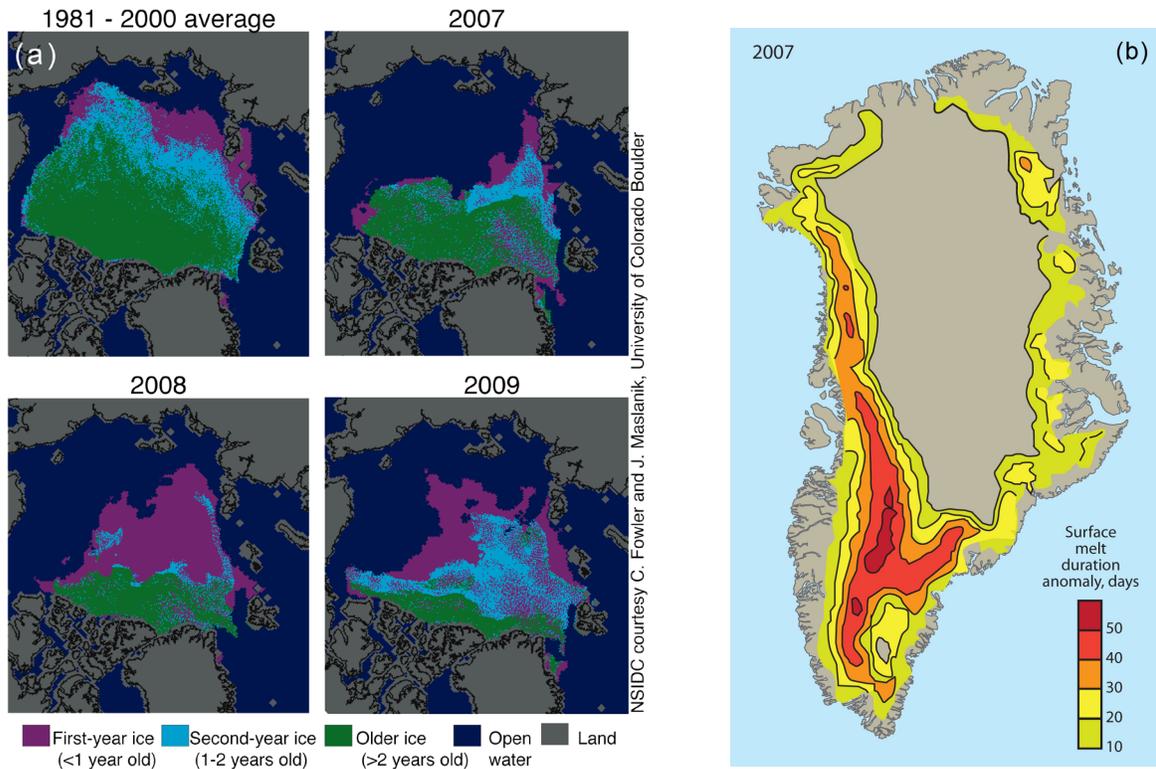


Figure 2-23. Evidence of Arctic Ice Melt. (a) Extent of summer Arctic sea ice for 2007-2009 compared to 1981-2000 average. (Source: National Snow and Ice Data Center, http://nsidc.org/news/press/20091005_minimumpr.html) (b) Duration of summer surface melt on Greenland in 2007 relative to 1973-2000 average (Source: Arctic Monitoring and Assessment Programme, 2009). Arctic summer sea ice has decreased by 40% since 1979, accompanied by an increasing discharge from the Greenland ice sheet. Natural variability may explain some of these changes, but the overall trend toward warming and melting has been attributed primarily to human-induced climate change (Min et al., 2008; Holland et al., 2008). Summer sea ice melt creates a feedback loop that amplifies warming as reflective white ice/snow surfaces are replaced by darker ocean waters, increasing sunlight absorption. Recent work suggests a link between Arctic sea-ice melt and increased glacier runoff in Greenland (Rennermalm et al., 2009).

and within the haze layer, while simultaneously contributing to surface dimming. Rather than a cooling effect from surface dimming, however, the atmospheric heating increases the downward longwave radiation and causes warming at the surface (Shaw and Stamnes, 1980; Quinn et al., 2008; Mauritsen et al., 2011). Any warming particle above a highly reflective surface can lead to heating of the entire surface-atmosphere aerosol column. In addition, the stable atmosphere above the Arctic prevents rapid heat exchange with the upper troposphere, increasing surface warming in the Arctic (Hansen and Nazarenko, 2004; Quinn et al., 2008).

Radiative forcing from both atmospheric concentration and deposition on the snow and ice has contributed to the surface temperature warming in the Arctic (Quinn et al., 2008). Simulations by

Flanner et al. (2007) suggest that the deposition of BC from sources in North America and Europe on Arctic sea ice may have resulted in a surface warming trend of as much as 0.5 to 1°C. Similarly, Shindell and Faluvegi (2009) found 0.5 to 1.4°C of warming from BC in the Arctic since 1890. For the BC snow albedo effect, Quinn et al. (2008) estimated a warming of 0.24 to 0.76°C, varying by season. Warming due to BC heating in the atmosphere is estimated to be a further 0.24°C in spring, 0.15°C in summer, and nearly zero in autumn and winter. In Table 2-6, we show estimates of temperature increases in the Arctic from various BC emissions sources (Shindell and Faluvegi, 2009; Flanner et al., 2007; Jacobson, 2010). Part of these increases in temperature may also have been “unmasked” in recent years from reductions in sulfate aerosols and its gaseous precursor, SO₂ (Shindell and Faluvegi, 2009). While sulfate aerosols have a negative

Table 2-6. Arctic Temperature Impacts from Emissions of BC from Different Sectors.

Scale of Physical Impact	Estimated Forcing or Temperature Change	Source of Emissions	Aerosols	Model	Reference	Notes
Arctic	0.5 – 1.4° C	100% of FF, BF, BB	BC, OC	GISS-ER	Shindell and Faluvegi (2009)	Aerosol indirect included “crudely.”
Arctic	0.5 – 1.6° C	100% of FF, BF, BB	BC, OC	NCAR-CAM3 and SNICAR	Flanner et al. (2007)	Range results from using fire frequencies in high year globally (1998) and low year (2001).
Arctic	1.2° C	100% of FF	BC, OC, minor inorganics	GATOR-GCMOM	Jacobson (2010)	
Arctic	1.7° C	100% of FF, BF	Same as above plus CO ₂ , CH ₄	Same as above	Jacobson (2010)	
Arctic	Indirect > 0	100% of All sources	All aerosols	Idealized calculations, observations	Mauritsen et al. (2010)	Indirect only, no direct or snow albedo effect.

FF = fossil fuel; BF = biofuel; BB = biomass burning.

radiative forcing, the reductions in sulfate aerosols have been strongly justified by improvements in air quality, acid rain, visibility, public health, and lessening of direct effects of sulfates on ecosystems. It has also been suggested that the potential cooling effects of BC—such as indirect radiative forcing and the ratio of BC to cooling components (e.g., OC)—may not be as important in the Arctic since the snow and ice albedo darkening is so dominant (Mauritsen et al., 2010).

While there are strong qualitative indications of Arctic snow and ice melt from BC, quantitative studies have only recently entered the peer-reviewed literature. Some studies have linked the local warming measured on the Greenland Ice Sheet to observations of a gradual loss of ice, and modeled the overall impact on the mass balance of the ice sheet. Box et al. (2004), for example, estimated the modeled ice sheet mass balance at -76 km³ per year, leading to a 0.24 mm sea level rise per year (contributing 15% of global sea level rise) during 1991-2000. Hanna et al. (2005) considered a longer time period, and estimated that the overall mass balance declined at a rate of -22 km³ per year in 1961-1990 and -36 km³ per year for 1998-2003, with melting during the past six years contributing 0.15 mm per year to global sea level rise. Finally, Thomas et al. (2006) reported accelerating mass loss between an earlier period (between 4 and 50 Gigatons (Gt) per year, depending on the model, from 1993-1999) and a more recent period (between 57 and 105 Gt per year, from 1999-2004). In a modeling study by Flanner et al. (2007), land snowmelt rates north of

50°N latitude (about 70 miles north of the U.S./Canada border in Minnesota) increased by 28% in 1998 and 19% in 2001 in the month preceding maximum melt when compared to control runs that did not include BC from large boreal fires that occurred in 1998 and 2001. Strack et al. (2007) found soot deposition in the Alaskan Arctic tundra created snow-free conditions five days earlier than model runs without BC deposition. Ongoing studies will help evaluate and constrain modeling simulations. Importantly, American, Norwegian, Russian, and Canadian research groups collaborated under the International Polar Year (2007-2008) program to survey BC concentrations in snow and ice north of 65°N latitude in both the Eastern and Western Arctic (Doherty et al., 2010).

The location of the emissions also matters for the magnitude of the effects in the Arctic, which has important implications for mitigation decisions. A recent study by the Arctic Monitoring and Assessment Program (Quinn et al., 2011) analyzed the radiative forcing impacts of BC emissions from different regions on the Arctic (direct and snow/ice albedo only). This study found that compared to the average emissions of BC from regions between 40°N and 50°N latitude, emissions of BC from between 50°N and 60°N latitude had about three times as much forcing impact in the Arctic on a per-ton basis. In addition, emissions from north of 60°N had seven times as much impact per ton. However, because total emissions are much larger in the southern regions, almost half of the total impact on Arctic forcing due to BC in this study was derived from

emissions from regions below 50°N, and most of the remainder was derived from emissions from regions between 50°N and 60°N.

BC emissions from near-Arctic countries have decreased since their peak in the early 20th century. This is supported by a downward trend in the observed concentrations of ambient and snow/ice BC in the Arctic (see section 5.6). However, BC deposited on areas covered permanently with ice and snow, such as the Greenland Ice Sheet, can remain in the ice sheet for hundreds of years, as seen in historical ice core records (Quinn et al., 2008). Although the sunlight that reaches the snow surface typically only penetrates between 10 and 20 cm deep, with the topmost 5 cm of snow receiving the most sunlight (Galbavy et al., 2007), as the Arctic warms and snow and ice melt, deeper, hidden BC that was deposited over decades may become exposed, enhancing the melting of snow and ice. The effect of BC on the snow and ice albedo in the Arctic thus can involve historical—in addition to present-day—BC deposition in the Arctic region.

An important uncontrolled source of BC in some near-Arctic countries is open biomass burning. Several recent studies have looked at the effect of these emissions on the Arctic. For example, Stohl et al. (2006) found that North American boreal forest fires lead to elevated concentrations of light absorbing aerosols including BC throughout the entire Arctic, with substantial implications for Arctic warming and enhanced snow albedo effects. Analyses conducted by Hegg et al. (2010) suggest that the dominant source of light-absorbing aerosols (including BC) in the Arctic region is biomass burning. Other studies have linked open biomass burning to reduced surface albedo and accelerated melting (Hegg et al., 2009; Generoso et al., 2007; Kim et al., 2005). Following agricultural fires in Eastern Europe in spring 2006, Stohl et al. (2007) measured record high air pollution levels and BC concentrations in parts of the Arctic above Europe. Similarly, in a series of studies, Warneke et al. (2009; 2010) found that spring fires in Russia (Siberia) and Kazakhstan can more than double the Arctic haze that builds up during the winter months.

2.6.5 BC Impacts in the Himalayas

The world's third largest snowpack after Antarctica and the Arctic is found in the Hindu Kush-Himalayan-Tibetan (HKHT) region. The mountain ranges that define this region fall primarily along the borders of Pakistan, Afghanistan, India, Nepal, and China (UNEP, 2008a). It is often referred to as the Earth's "third pole." Atmospheric warming associated with BC is believed to be a significant factor in the observed

increases in melting rates of glaciers and snowpack in the HKHT (Barnett et al., 2005; Lau and et al., 2010; UNEP, 2008a; Thompson, 2010). Ramanathan and Carmichael (2008) and Ramanathan et al. (2007) suggested that the advection of air warmed by BC over the Himalayas has played a role comparable to that of GHGs in the observed retreat of Himalayan glaciers. A recent study by Carmichael et al. (2009) also shows that BC throughout Asia has an atmospheric warming potential of about 55% of that attributed to CO₂.

High radiative forcing estimates have been calculated for the Himalayas due to the large amount of mountain snow and ice cover as well as the proximity to high emissions of BC from parts of China and the Indian subcontinent. Flanner et al. (2007) calculated an average forcing in this region of +1.5 W m⁻² with short-term forcing of up to 20 W m⁻² in the spring. Translating this to temperature, Flanner et al. (2009) attributed an increase in the land-averaged March-May surface temperature in Eurasia of 0.93°C from BC and organic matter in the atmosphere and deposited on the snow.

BC can alter snowpack and glacier extent and retreat through two mechanisms, the first being increasing and decreasing precipitation as discussed in section 2.6.1.4, and the second being local warming, especially through deposition, increasing the rate of melt. Lau et al. (2010) found that heating of the atmosphere by dust and BC leads to widespread enhanced warming over the Tibetan Plateau and accelerated snowmelt in the western Tibetan Plateau and Himalayas. Menon et al. (2010) show observed trends in snow cover in the Himalayas, with a spatially heterogeneous pattern of decreases and increases of up to 17% from 1990 to 2001, where the area of decreases is much larger than the area of increases. Menon et al. (2010) simulated similar heterogeneous snow cover changes due to aerosol emissions, showing that the influence of the aerosols was larger than the influence of changing sea surface temperatures over that time period. Over Eurasia, Flanner et al. (2009) conducted a modeling study that found the combination of strong snow albedo feedback and large fossil fuel and biofuel emissions of BC and organic matter from Asia induce 95% as much springtime snow cover loss as anthropogenic CO₂ alone. The effects on glaciers are not well quantified, but Xu et al. (2009a) found evidence that soot deposited on Tibetan glaciers has been a significant contributing factor to observed rapid glacier retreat. Changes in the timing and extent of melting may adversely affect regional freshwater resources in this region, which relies heavily on this melt (Carmichael et al., 2009).

2.6.6 Summary of BC Impacts in Key Regions

As described in the previous sections, the climate-related effects of BC can vary considerably across regions. Table 2-7 provides an overview of the regional variability in terms of BC's effects on radiative forcing, temperature, precipitation, and snow and ice across the United States, Asia, and the Arctic. In addition, Figures 2-13 and 2-15 are useful for understanding the regional variability of BC's radiative forcing effects.

2.7 Metrics for Comparing Black Carbon Impacts to Impacts of Other Climate Forcers

In response to Congress's request for an assessment of potential comparative metrics, this section summarizes a number of different approaches to comparing the effects of BC to CO₂ and other GHGs, but cautions that there is no one "best" metric; rather, the utility of a metric depends on the desired environmental outcome and policy objective. Therefore, this section begins by introducing the concept of using metrics for comparing BC-related

Table 2-7. Climate Effects of BC in the United States, Asia, and the Arctic (Summary).

Effects	U.S.	Asia	Arctic
Radiative Forcing Effects	<ul style="list-style-type: none"> Estimates of direct radiative forcing of BC over the United States range from 0.1 to 0.7 W m⁻². 	<ul style="list-style-type: none"> South and East Asia have some of the world's highest estimates of radiative forcing, but large ABCs exert a counterbalancing dimming effect at the surface. Average annual snow and ice albedo forcing in the Tibetan Plateau has been estimated to be 1.5 W m⁻², with local instantaneous forcing up to 20 W m⁻². 	<ul style="list-style-type: none"> Springtime Arctic forcing has been estimated to be 1.2 W m⁻² (direct) and 0.53 W m⁻² (snow albedo).
Temperature Effects	<ul style="list-style-type: none"> No studies were identified for U.S. temperature effects from BC. All global modeling studies include the temperature effects over the U.S., but results are difficult to extract. Estimates of average warming from BC in the Northern Hemisphere range from 0.29°C to 0.54°C. 	<ul style="list-style-type: none"> Over the Himalayan region, atmospheric BC was estimated to result in up to 0.6°C of warming. 	<ul style="list-style-type: none"> BC deposited on snow results in warming of roughly 0.4 to 0.5°C, varying by season. Atmospheric BC was estimated to contribute roughly 0.2°C in spring, 0.1°C in summer, and nearly zero in autumn and winter.
Precipitation Effects	<ul style="list-style-type: none"> One study found little change in the amount of precipitation in the western United States as a result of BC effects. Other studies have found that rainfall patterns in the eastern United States match PM emissions, but not specifically those of BC. 	<ul style="list-style-type: none"> The cooling at the surface leads to reduced evaporation and precipitation as well as changes in sea-land temperature gradients. Precipitation and temperature gradient modifications can lead to shifts of regional circulation patterns such as a decrease in the Indian and Southeast Asian summer monsoon rainfall and a north-south shift in eastern China rainfall. 	<ul style="list-style-type: none"> No studies were identified for Arctic precipitation effects.
Snow and Ice Effects	<ul style="list-style-type: none"> In the western United States, BC deposition on mountain glaciers and snow produces a positive snow and ice albedo effect, contributing to the snowmelt earlier in the spring. Early snowmelt reduces the amount of water resources that normally would be available later in the spring and summer, and may contribute to seasonal droughts. 	<ul style="list-style-type: none"> BC atmospheric warming is believed to be a significant factor in the melting of the HKHT glaciers and snowpack. The deposition of BC on glaciers and snowpack in Asia also has a strong snow and ice albedo positive feedback that accelerates melting of the glaciers and snow, with implications for freshwater availability and seasonal droughts. 	<ul style="list-style-type: none"> BC may increase snowmelt rates north of 50°N latitude by as much as 19-28%. Soot deposition in the Alaskan Arctic tundra created snow free conditions five days earlier than model runs without BC deposition.

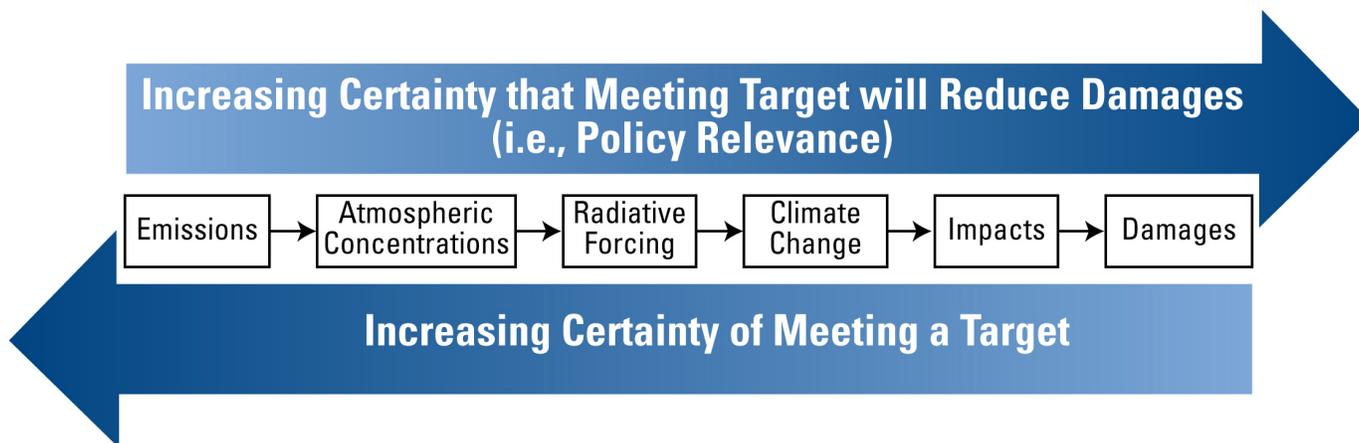


Figure 2-24. Cause and Effect Chain from Emissions to Climate Change, Impacts, and Damages. (Adapted from Fuglestedt et al., 2003.) The arrows indicate that a policy could focus on different elements along the causal chain and, depending on whether the policy focuses on the emissions or damages end of the chain, can determine the certainty of meeting the stated policy target versus the certainty of reducing damages at issue.

impacts to those of other climate forcers. It explains some of the approaches to developing metrics and provides a comparison of common metrics used for GHGs and for BC. This section concludes with a discussion of the most salient limitations associated with specific metrics and with using metrics in general.

The goal of a *metric*, as used in this report, is to quantify the impact of a pollutant relative to a common baseline. Such metrics can be used to compare between two or more climate forcers (e.g., CO₂ versus CH₄), or to estimate the climate effects of different emissions sources (or mitigation measures). Metrics that enable comparisons among pollutants or sources based on common denominators can also be used for the implementation of comprehensive and cost-effective policies in a decentralized manner (e.g., in a market-based climate program) so that multi-pollutant emitters can compose mitigation strategies (Forster et al., 2007).

Climate metrics are often defined relative to a baseline pollutant (usually CO₂) and focus on a particular climate impact (such as radiative forcing or temperature) that would be altered due to a change in emissions. For example, in EPA's annual *Inventory of U.S. Greenhouse Gas Emissions and Sinks*, the GWP metric is used to convert all GHGs into "CO₂-equivalent" units. Importantly, metrics such as GWP have been used as an exchange rate in multi-pollutant emissions policies and frameworks (IPCC, 2009). The key assumption when developing a metric is that two or more climate forcers are

comparable or exchangeable given the policy goal. That is to say, one pound of apples may be comparable to or exchangeable with one pound of oranges if the goal is not to overload a truck, but not if the goal is to make apple cider (Fuglestedt et al., 2010). Therefore, when used as an exchange rate in multi-pollutant emissions framework, a metric allows substitution between climate forcers which are presumed to be equivalent for the policy goals (Forster et al., 2007).

Metrics can also be used to prioritize among mitigation measures designed to control emissions of similar compounds from different sources. As described previously in this chapter, aerosols are composed of numerous components, and these different components can contribute to both warming (BC) and cooling. A metric can aggregate these effects in order to determine the relative contribution of a given source or measure.

2.7.1 Metrics Along the Cause and Effect Chain

For both BC and GHGs, there is a cause and effect chain starting with anthropogenic emissions and leading to changes in concentrations, radiative forcing, physical climatic changes, and impacts on human and natural systems (Figure 2-24). Some of the links in this cause and effect chain may be simultaneous rather than sequential. For example, the atmospheric loading of aerosols affects dimming and precipitation directly, rather than mediated through radiatively induced temperature

changes. Nor is the chain always unidirectional. Climatic changes can lead to changes in atmospheric concentrations of climate-forcing pollutants (e.g., changes in precipitation will change aerosol lifetimes) or even emissions of those pollutants (e.g., changes in temperature affect fossil fuel consumption for heating and cooling needs, which affects emissions of particles and precursors). There are uncertainties at each stage of the cause and effect chain, and these uncertainties compound over multiple steps of the chain. The uncertainties for BC are generally larger at all stages of the causal chain compared to the long-lived GHGs (for reasons discussed in this and other chapters of this report).

Within the climate change field, metrics have been calculated for changes in radiative forcing, global mean temperature, and monetized damages. The closer the metric is to the emissions end of the chain, the less uncertainty there is in how to calculate the metric; it is easier to determine how a change in emissions will change concentrations than it is to determine how a change in emissions will change temperature (a calculation which requires several intermediate steps). Additionally, the further along the chain, the more physical systems (and economic systems) need to be included in order to calculate the metric. However, if a reduction in damages is considered the ultimate objective of the policy, then a metric that focuses explicitly on impacts or damages best represents that objective. Since the economic value of damages (expressed in dollars) is one of the easiest metrics for the public and policymakers to place in context, there has been a great deal of interest recently in calculating the monetary value of climate change impacts associated with different pollutants (see Chapter 6). The choice

of a metric can be considered in part a choice about how to allocate uncertainty between calculation of the metric and the representativeness of the metric for the ultimate impacts of interest.

Fuglestedt (2009) identified the following considerations for developing a metric for climate forcers (see Table 2-8 for examples of how commonly used metrics address these considerations):

1. What climate impact is of interest for the policy being considered?
2. What climate forcer will be used as the baseline for comparison?
3. What is the temporal frame for emissions? Is it an instantaneous pulse or a sustained change in emissions?
4. What is the temporal frame for the impact? 10 years, 50 years, 100 years? Is the impact considered only at the end point of the time frame, or integrated over the period?
5. Does the metric address the magnitude of change or the rate of change or both?
6. What is the spatial dimension of the metric for both emissions and impacts? Is it global or regional?
7. What economic considerations should be taken into account? How are damages in the far future weighed compared to damages in the near term?

Table 2-8. Examples of Commonly Used Metrics for GHGs.

Metric Type	Climate Impact	Baseline Forcer	Emissions Type	Spatial Scale	Includes Rate of Change?
GWP (Global Warming Potential)	Integrated radiative forcing	CO ₂	Pulse	Global	No
GTP-pulse (Global Temperature Potential)	Temperature	CO ₂	Pulse	Global	No
GTP-sustained	Temperature	CO ₂	Sustained	Global	No
STRE (Surface Temperature Response per unit continuous Emission)	Temperature	CO ₂	Sustained	Global	No
SFP (Specific Forcing Pulse)	Energy	Joules/gram	Pulse	Global or regional	No
Cost-effectiveness Metrics (e.g., Manne and Richels, 2001, Global Cost Potential)	Mainly temperature	CO ₂ or \$ value	Optimal emissions calculation	Global	Optional
Value of Damages (e.g., Social Cost of Carbon, Global Damage Potential)	Range of climate damages	\$ value	Pulse	Global	Limited

First, the climate impact must be identified because the effectiveness of a given metric is dependent on the primary policy goal. Considerations 2 through 7 are then framed by the selected climate impact. This is important because choosing an inappropriate metric could lead to policy decisions that ultimately result in undesirable climate or economic impacts.

2.7.2 Commonly-Used Metrics for GHGs

Article 2 of the United Nations Framework Convention on Climate Change (UNFCCC) calls for a policy that addresses the magnitude and the rate of climate change as well as the cost effectiveness of controlling emissions (IPCC, 2009). Therefore, appropriate metrics could cover either the physical or economic dimensions of climate change, or both. A number of metrics have been developed and refined for application to CO₂ and other long-lived GHGs. These metrics are summarized in Table 2-8 and described further below. Their potential applicability to BC is considered in the next section. Note that two of the metrics listed in the table (SFP and STRE) were developed specifically for application to short-lived climate forcers like BC, and are discussed only in section 2.7.3.4 below.

Five considerations are listed in Table 2-8. The first, climate impact, refers to where the metric falls on the cause-effect chain shown in Figure 2-24. The second, baseline forcer, lists whether the metric is measured in comparison to CO₂, or in absolute units (whether dollars or energy). The third column notes what kind of emissions change is being considered. A “pulse” of emissions refers to an effectively instantaneous release of that pollutant (though sometimes that release is considered to be spread out over a year). A pulse analysis is appropriate for a one-time trading of emissions permits, but may not be as realistic for analyzing investment decisions which spread reductions out over time (though a longer term reduction can be approximated as a series of pulses). Therefore, other analyses consider the possibility that an emission reduction (or increase) will be permanent (i.e., sustained over time). The third temporal option is to calculate the optimal emissions path, which is discussed in more detail in section 2.7.2.3 (cost-effectiveness metrics). The fourth column shows that most metrics have been designed to be used on a global scale, though some of these might be adaptable for regional impacts. Finally, most metrics consider temperature change or damages either at a single point in time or summed over time: only a few consider that there may be value in limiting the rate of change in addition to reducing the absolute magnitude of the change.

Table 2-8 is also ordered in a rough approximation of the transparency of the metric. Metrics which are transparent and easy to calculate are likely to be more readily accepted for policy use than those which require complex modeling. The GWP is in widespread use and can be calculated based only on knowing the average lifetime of a molecule of a gas and the radiative forcing caused by that molecule. The remaining metrics require the use of computer models of more or less complexity in order to calculate, and if the metric is sensitive to assumptions involved in the modeling then that could have potential for controversy.

2.7.2.1 Global Warming Potential

To date, the most widely established and well-defined metric is the GWP. The definition of the GWP by the IPCC (2007) is

“An index, based upon radiative properties of well-mixed greenhouse gases, measuring the radiative forcing of a unit mass of a given well-mixed greenhouse gas in the present-day atmosphere integrated over a chosen time horizon, relative to that of carbon dioxide. The GWP represents the combined effect of the differing times these gases remain in the atmosphere and their relative effectiveness in absorbing outgoing thermal infrared radiation. The Kyoto Protocol is based on GWPs from pulse emissions over a 100-year time frame.”

The identified climate impact the GWP addresses is globally averaged change in radiative forcing and its baseline climate forcer is CO₂ (e.g., the GWP of CO₂ is defined to be 1).¹⁵ The temporal frame for emissions is a pulse. The GWP provides the magnitude, but not the rate of change, of the integrated radiative forcing over a given time frame. The time frame is usually 100 years, but 20-year and 500-year GWPs are sometimes presented to show how GWPs would differ if short-term or long-term impacts are given more weight. Finally, the GWP, which addresses only radiative forcing, a physical metric, does not take into account any economic dimension.

As discussed below, there have been a number of criticisms of the GWP in the peer-reviewed literature (e.g., O’Neill, 2000; Shine, 2009), mainly focused on either the inability of the GWP to capture key differences between gases (such as different lifetimes) or the failure of the GWP to incorporate economic considerations. Despite such criticisms, at the time of the Kyoto Protocol in 1997, the GWP was

¹⁵ The GWP is calculated as the ratio of the Absolute Global Warming Potential (AGWP) of a given gas to the AGWP of CO₂. The AGWP has units of W m⁻²yr g⁻¹.

adopted as the metric used in climate negotiation. While acknowledging that there are shortcomings involved in using GWPs even for comparisons among the long-lived gases, a recent IPCC Expert Meeting on the topic found that GWPs were still a useful measure for these gases (IPCC, 2009). It remains the most accepted metric due to its simplicity, the small number of input parameters, the relative ease of the calculation, and a lower level of uncertainty compared to some alternatives (Shine et al., 2005). The GWPs as calculated by the IPCC Second Assessment Report (Schimel et al., 1996) currently remain the standard GWPs used for the official U.S. GHG emissions inventory compiled annually by EPA, as required by UNFCCC reporting guidelines.¹⁶

2.7.2.2 Global Temperature Potential

One alternative metric that has received recent attention is the GTP. Like the GWP, the GTP is a physical metric. Whereas the GWP considers change in globally averaged radiative forcing, the GTP compares the globally averaged temperature change at a given point in time resulting from the emission of two climate forcers of equal mass (Shine et al., 2005). The GTP moves one step further down the cause and effect chain and addresses a climate response to radiative forcing, the global-mean surface temperature change. The GTP therefore includes more physical processes, such as the heat exchange between the atmosphere and ocean, than the GWP. This introduces more uncertainty to the metric, and can require the use of more complex models in order to calculate the GTP value. In addition, while the GWP represents the integrated radiative forcing of a pulse of emission over a given time period, the GTP is evaluated at a given point in time (IPCC, 2009). Like GWPs, the GTP can be calculated over a variety of timescales, with 20, 100, and 500 years being the timescales most commonly presented. There are advantages and disadvantages to using either the GWP or a GTP, and they may each address different policy goals and may be more relevant to different climate forcers and time frames, depending upon the policy need. To date, however, the GTP has not been used as a metric for trading gases in international, national, or regional accords.

There are two versions of the GTP: one that involves the effects of a pulse of emissions, and another that involves a sustained reduction of emissions. The latter version of the GTP results in comparative values between different gases that are similar to the values calculated using GWPs. The former version of the GTP, by contrast, leads to longer-lived

gases being given more relative weight because a pulse of a short-lived gas has very little impact on temperatures many years in the future.

The GTP can also be calculated as a function of a future global temperature stabilization target. One criticism of a number of metrics is that they are not compatible with a goal of stabilization because the target is not part of the metric. Manne and Richels (2001) developed a methodology to calculate a time-dependent metric (referenced below as a cost-effectiveness metric) that would change as a target level was approached. For example, if the target is not to exceed a 2 degree temperature change above preindustrial, then when global temperatures are still only 1 degree above preindustrial, and therefore the target temperature is still decades away, the metric will place weight on long-lived gases like CO₂. But as the target temperature is approached, the time to reach that target becomes short, and the metric places weight on the strong, short-lived forcers like CH₄ and BC.

Shine et al. (2007) used a similar approach to develop a time-dependent GTP, the GTP(t). Shine et al. applied the GTP(t) to BC using a target of 2°C, and found that for a low emissions scenario, GTP(t) starts at about 2 in 2010, rising to 1,000 by 2080. But for a high emissions scenario, GTP(t) can start at 200, reach 1,000 in 2030, and reach 20,000 in about 2045. While this approach is one of the few approaches that are truly compatible with a stabilization target, there are some drawbacks. Drawbacks include the dependence on assumptions about future emissions scenarios, the undefined nature of the metric *after* reaching the stabilization target, and the dependence of the metric on computer modeling, which reduces transparency. In addition, policymakers might not desire a metric whose value can change by orders of magnitude over several decades and without a transparent and predictable schedule. One advantage of the GTP(t) and related metrics is that they can easily be adapted to include a rate of change goal; for example, rather than just constraining the metric to reach a 2°C target, it is also possible to value the rate of change as well by adding on a constraint that the temperature not increase more than a given amount in any given decade. Such an additional constraint would increase the value of short-lived substances like BC.

2.7.2.3 Cost-Effectiveness Metrics

Manne and Richels (2001) examined relative tradeoffs between different gases that vary over time and are calculated to optimally achieve a given target using a computer model that included economic considerations. Similarly, the Global Cost

¹⁶ See <http://www.epa.gov/climatechange/emissions/usgginventory.html>.

Potential (GCP), compares the relative marginal abatement costs for two climate forcers when a given climate change target is achieved at least cost (IPCC, 2009). These approaches define a temperature or radiative forcing target and calculate the relative (or absolute) dollar value that should be imposed on different gases in order not to exceed that target.

2.7.2.4 GHG Metrics for Measuring Economic Impacts

Two metrics, the Global Damage Potential (GDP) and the social cost of a pollutant, involve monetization of the damages of climate change (see detailed discussion in Chapter 6). The GDP compares the relative damage resulting from an equal mass of emissions of two climate forcers (IPCC, 2009). The social cost calculation has most commonly been used for CO₂ alone, where it is referred to as the Social Cost of Carbon (SCC). However, even where risks and impacts can be identified and quantified with physical metrics, it may be difficult to monetize these risks and impacts (e.g., such as ecosystem damage or the potential to increase the probability of an extreme weather event) such that an accurate cost-benefit comparison could be undertaken. Both the GDP and the social cost calculation depend on the physical aspects of the climate system as well as the economic linkages between climate change impacts and the economy (IPCC, 2009). Therefore, the GDP and the social cost require calculations of the entire cause and effect chain, but as a result contain a large amount of uncertainty.

2.7.3 Applicability of Climate Metrics to BC

This section discusses the use of well-established metrics such as the GWP and GTP as they relate to BC emissions and identifies alternative metrics that may be more relevant to BC. As discussed earlier in this chapter, BC influences the climate differently than the warming effects of GHGs. These differences have important implications for identifying appropriate metrics to compare climate impacts (and reductions thereof). Table 2-1 compared some of BC's climate attributes and effects to those of CO₂. The implications of these differences with respect to metrics are discussed here.

As described in detail below, the significant differences between BC and CO₂ make applying the metrics introduced in the previous section difficult and, for some purposes, inappropriate. One of the most essential factors to consider is that BC is most clearly related to short-term climate impacts, and is principally a regional pollutant. The lifetime of BC (weeks) is much shorter than the mixing time of the atmosphere (1 to 2 years), so the climate impacts of

BC depend heavily on where and when it is emitted. In comparison, the shortest-lived GHG in the Kyoto basket has a lifetime longer than one year, and the majority of the Kyoto gases have lifetimes ranging from decades to millennia. In addition, the variation in atmospheric concentrations of BC among regions contrasts with the well-mixed nature of most GHGs. This distinction has not been captured in most metrics to date. Thus, focusing on long-term, global average radiative forcing impacts—the frame of reference for long-lived GHGs—may lead to distorted policy decisions about BC. Conversely, focusing on short-term or regional impacts may be inappropriate for decisions involving long-lived GHGs. The following sections discuss how different physical (GWP, GTP, SFP, and STRE) and economic metrics have been used to compare BC to other substances.

2.7.3.1 Global Warming Potential

While a GWP can be calculated for BC, there are reasons that GWPs may be less applicable for this purpose due to the different nature of BC compared to GHGs, in terms of various physical properties and the fact that unlike GHGs, BC is not well mixed in the atmosphere. However, because GWPs are the most commonly used, and only official, metric in climate policy discussions, many studies have calculated GWPs for BC. One-hundred-year GWPs for BC in the literature range from 330 to 2,240. That is to say, 330 to 2,240 tons of CO₂ would be required to produce the same integrated radiative effect over 100 years as one ton of BC. Some of the factors that account for the range in these estimates include the use of different and uncertain indirect and snow/ice albedo effects estimates, use of a different CO₂ lifetime for the baseline, and recognition of the dependence of a GWP for BC on emissions location.

Using time periods shorter than 100 years has also been explored for determining the GWPs of BC. Those who are concerned with near-term impacts (such as Arctic ice retreat) sometimes suggest 20-year GWPs as more appropriate for short-lived forcers such as BC (CATF, 2009b). Jacobson (2007) estimates a 20-year GWP for BC of 4,470. However, for those concerned about the long-term problems of climate change, even 100-year GWPs may be considered too short (IPCC, 2009). Because BC is a short-lived species, the shorter the policy-relevant time horizon considered, the greater the relative importance of BC compared to CO₂ (and vice versa: the longer the relevant time horizon, the less important BC is compared to CO₂). If the focus is on achieving immediate climate benefits within a 10- to 20-year time period, the 20-year GWP provides a more realistic picture of the impact of reductions

in different species in the near-term. On the other hand, if the concern is to identify measures that will help avert climate change at a broad scale, over a longer time frame, as the problem is generally conceptualized, a 20-year time horizon is insufficient, and the 100-year GWP is a more relevant metric.

2.7.3.2 Global Temperature Potential

GTPs, as described previously, evaluate the impact on temperature at a given time. Studies have applied the GTP using approaches that differ with respect to how the emissions are reduced and how the impacts are calculated. Boucher and Reddy (2008) use a short, pulse-like (1-year) reduction of emissions and find that the 100-year GTPs are about a factor of 7 smaller than the corresponding GWPs. Berntsen et al. (2006) reduced BC emissions for a 20-year time span (approximately the lifetime of a given investment in abatement technology) and found that the 100-year GTP of BC was about 120 to 230 (i.e., reducing 120 to 230 tons of CO₂ has the same impact on temperatures in 100 years as reducing 1 ton of BC).

Several papers have recently summarized different BC GWP and GTP estimates (Sarofim, 2010; California Air Resources Board, 2010; Fuglestvedt et al., 2010). However, of the studies surveyed by these three papers, only Hansen et al. (2007a) considered indirect cloud interactions of BC and only a few included estimates for metrics of co-emitted OC. If

co-emissions are not included, then any metric will likely overestimate the globally averaged climate benefits of reducing BC. Inclusion of indirect effects could either increase or decrease the calculated value of the metric.

Figure 2-25, based on Fuglestvedt et al. (2010), summarizes a number of studies that attempted to develop metrics for comparing CO₂ and BC. This figure shows how the GWP metric depends on the time horizon used (20 years, 100 years, and 500 years). Additionally, for the first four studies, the range of values results from a dependence of the GWP on the region in which the emission occurs. The difference between the studies is the result of differences in the climate models used to link the emissions to the temperature change. Figure 2-26 shows a similar analysis from Fuglestvedt et al. (2010) which evaluates the equivalent GTP for these different models.

Fuglestvedt et al. (2010) show that the metric for comparing BC to CO₂ can range from a ton of BC being equivalent to 48 tons of CO₂ based on a 100-year GTP (which measures the temperature change 100 years after a pulse of emissions) in one model, to 4,900 tons of CO₂ based on a 20-year GWP (which integrates the total radiative forcing impact of a pulse of emissions over the 20-year time span) in another model. The variation between GWPs or GTPs for emissions from different

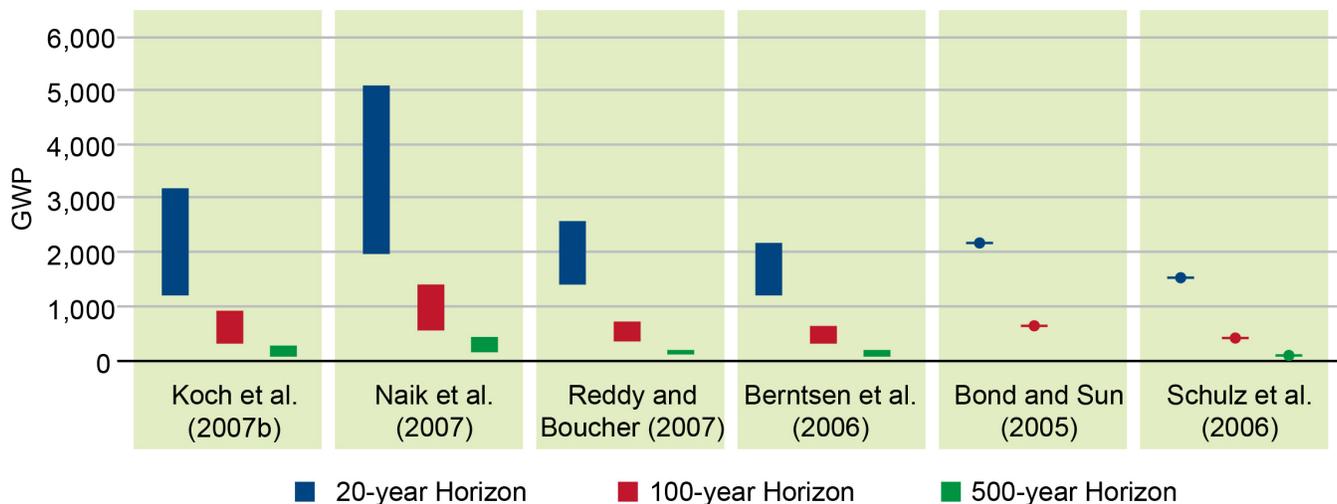


Figure 2-25. Ranges and Point Estimates for Regional Estimates of GWP Values for One-Year Pulse Emissions of BC for Different Time Horizons. The GWP values in the Y axis of the figure refer to the number of tons of CO₂ emissions which are calculated to be equivalent to one ton of BC emissions based on the particular metric. (Adapted from Fuglestvedt et al., 2010.) Note that the first four studies referenced evaluated GWP values for different sets of regions; Bond and Sun (2005) and Schulz et al. (2006) produced global estimates only.

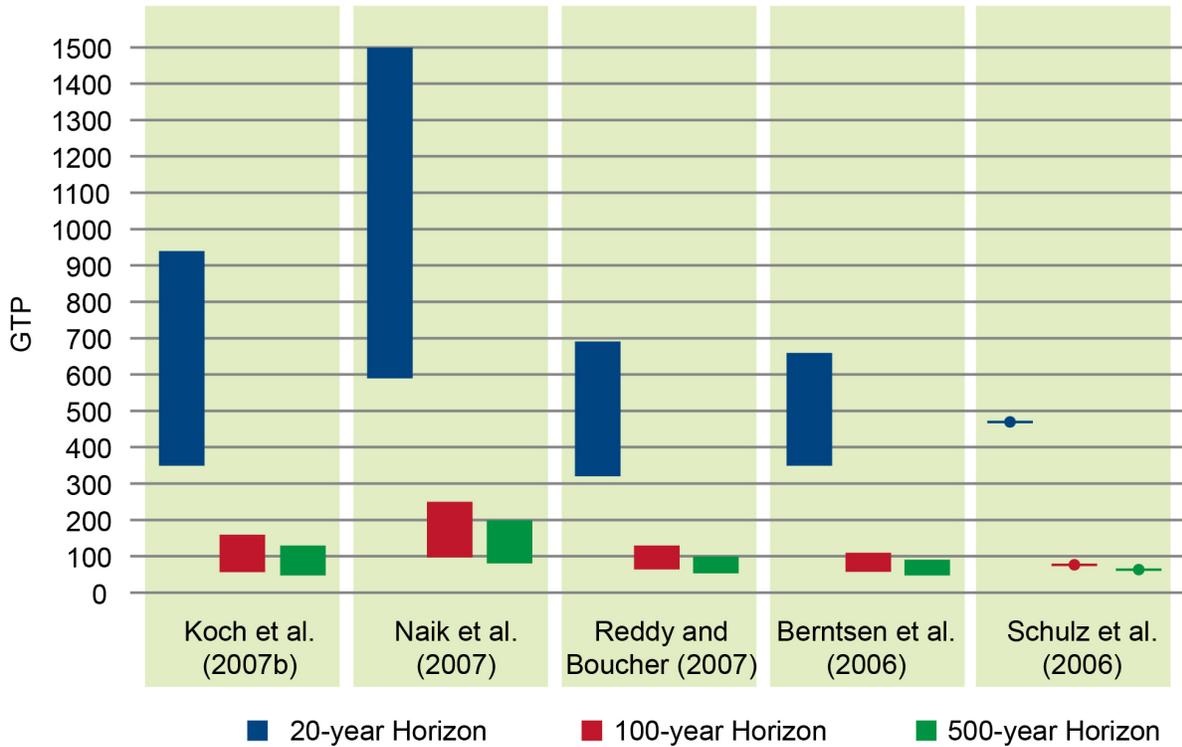


Figure 2-26. Ranges and Point Estimates for Regional Estimates of GTP Values for One-Year Pulse Emissions of BC for Different Time Horizons. The GTP values in the Y axis of the figure refer to the number of tons of CO₂ emissions which are calculated to be equivalent to one ton of BC emissions based on the particular metric. (Adapted from Fuglestvedt et al., 2010) Note that the five studies referenced evaluated radiative forcing estimates for different sets of regions (which were translated into GTP values by Fuglestvedt et al.); Schulz et al. (2006) produced global estimates only.

locations demonstrates how variability in convective properties, exposure to sunlight, and different surface albedos can cause the effect of a given unit of emissions of BC to vary. Given a specific timescale, metric, and computer model, the two figures show that this dependence on emissions location can lead to changes in GWP or GTP by up to factor of three. Such dependence on emissions location for long-lived GHGs does not come into play when calculating their GWPs.

Sarofim (2010) also summarized a number of studies, and further analyzed how the GWP estimate depended on inclusion of either fossil fuel OC co-emissions or snow albedo impacts. Sarofim (2010) found that inclusion of these processes can change the value of the metric by about a factor of two. Other effects that were not quantified in the paper, but that can lead to significant differences between model estimates of GWPs, are the inclusion of indirect effects on clouds and the assessment of a larger range of sectors and co-emission types. Additionally, because most metrics use CO₂ as a

baseline forcer, the use of different carbon cycle models can significantly influence the metric values for BC. Some researchers may report metric values in carbon equivalents, rather than CO₂ equivalents, which leads to a factor of 3.7 difference.

2.7.3.3 Specific Forcing Pulse

The SFP is a relatively new metric proposed by Bond et al. (2011) to quantify climate warming or cooling from short-lived substances (i.e., substances with lifetimes of less than four months). This metric is based on the amount of energy added to the Earth system by a given mass of the pollutant. The rationale for developing this new metric was that short-lived substances contribute energy on timescales that are short compared to time scales of mitigation efforts, and therefore can be considered to be "pulses." Bond et al. (2011) find that the SFP of the direct effect of BC is 1.03±0.52 GJ/g, and with the snow albedo effect included is 1.15±0.53 GJ/g. They also find that the SFP for OC is -0.064 (from -0.02 to -0.13) GJ/g, which leads to a conclusion that

for direct forcing only, a ratio of about 15:1 for OC to BC is close to climate neutral. However, this does not include cloud indirect effects or co-emissions of substances other than OC. Bond et al. also find that the SFP varies by 45% depending on where the BC is emitted. While the paper notes that fundamental differences in temporal and spatial scales raise concerns about equating the impacts of GHGs and short-lived aerosols, they do use the SFP to calculate a GWP for the direct effect of BC of 740 (± 370), for both the direct and the snow albedo effect of BC of 830 (± 440), and for organic matter of -46 (from -18 to -92).

This metric is mathematically similar to the Absolute Global Warming potential (see footnote in GWP section), but is applied somewhat differently. Additionally, the use of this metric for regional impacts is interesting, though as discussed earlier, the regional pattern of radiative forcing (or energy input) is not necessarily the same as the regional pattern of temperature response to that forcing.

2.7.3.4 Surface Temperature Response per Unit Continuous Emission

Another new metric, the STRE has been proposed by Jacobson (2010). The STRE is similar to the sustained version of the GTP. Jacobson found that the STRE (which he compares to GWPs) for BC on the 100 year time scale is 2,900 to 4,600 for BC in fossil fuel soot and 1,060 to 2,020 for BC in solid-biofuel soot. The uncertainty ranges presented by Jacobson depend on his assumption that CO₂ will decay exponentially with either a 30- or a 50-year lifetime. The use of a more sophisticated carbon cycle model or the Bern carbon cycle approximation from the IPCC (which is a sum of 4 exponentials rather than a single exponential as in the Jacobson calculations) would result in a lower STRE and would be more comparable with other approaches. Jacobson also presents estimates of the combined BC plus OC STRE, finding that the STRE for emissions of BC plus OC from fossil fuel soot ranges from 1,200 to 1,900 and for emissions from biofuel soot the STRE ranges from 190 to 360.

2.7.3.5 Economic Valuation Metrics

Economic valuation approaches for BC that focus on valuing climate damages from a comprehensive, societal standpoint are discussed in detail in Chapter 6. For reasons discussed in that chapter, techniques used to value the climate damages associated with long-lived GHGs are not directly transferrable to BC or other short-lived forcers. In fact, most such approaches have focused exclusively on valuing the climate impacts of CO₂, and may not even be

transferrable to other GHGs. Additional work is needed to design approaches for valuing the climate impacts of BC directly, and to incorporate those approaches into metrics comparable to the SCC.

2.7.4 Using Metrics in the Context of Climate Policy Decisions

There is currently no single metric widely accepted by the research and policy community for comparing BC and long-lived GHGs. In fact, some question whether and when such comparisons are useful. For example, there are concerns that some such comparisons may not capture the different weights placed on near-term and long-term climate change. However, there are multiple reasons to compare BC to other short-lived and long-lived climate substances, including offsets, credit trading, evaluation of net effects of a mitigation option, or illustrative analyses.

The choice of a metric depends greatly on the policy goal. No single metric will accurately address all the consequences of emissions of all the different climate forcers, and all of the differences between BC and the well-mixed gases must be considered. The appropriate metric to use depends on a range of factors, such as: the time scale (20 years, 100 years, or more), the nature of the impact (radiative forcing, temperature, or more holistic damages), concern over different processes (indirect effects, snow albedo changes, co-emissions), and whether sources and impacts should be calculated regionally or globally. It is important to note that different climate models will yield different results even if the same metric definition is chosen. Taking several of these factors into account, especially the use of different time scales, a ton of BC has been calculated to be equivalent to anywhere from 48 tons of CO₂ to 4,600 tons of CO₂. For comparison, the UNEP/WMO assessment, looking only at the 100 year timescale, estimated that BC could be 100 to 2,000 times as potent as CO₂ per ton (UNEP and WMO, 2011a). Certainly, the appropriateness of the comparison depends on the policy question at hand, and the differences in lifetime, uncertainties, co-emissions, modes of interaction with the climate system, and non-climatic effects such as human health should be evaluated when choosing a metric. This section highlights how these differences affect the metric choice.

The tradeoff between capturing short-term and long-term impacts is not strictly a scientific consideration but also a policy question. Much like the original choice of 100 years for the GWP was a policy compromise between long-term and short-term impacts; policymakers may consider

whether using a GWP or GTP metric is an acceptable compromise given a desire to compare BC and the long-lived GHGs. A key question is how the metric is used to inform the policy decision. The NRC has warned against delaying CO₂ reductions in favor of short-lived forcer mitigation, suggesting that CO₂ emissions control and control of short-term forcing agents could be thought of instead as “two separate control knobs that affect entirely distinct aspects of the Earth’s climate” (National Research Council, 2011). The results of the UNEP/WMO assessment suggest that the two strategies are complementary and should be pursued simultaneously, with BC reductions forming part of a larger strategy for near-term climate change and CO₂ programs influencing climate over the longer term (UNEP and WMO, 2011a). Such an approach could incorporate separate metrics for short-lived and long-lived species. One metric would be appropriate for guiding global emissions of climate forcers to achieve stabilization of GHG concentrations in the long-term, while another metric would focus on mitigating near-term warming and could be used to guide regional emissions reductions in short-lived climate forcers to reduce the impacts on regional forcing, precipitation, and ice/snow melt. It is important to recognize that long-term stabilization of CO₂ concentrations requires limiting total cumulative emissions of CO₂ and that CO₂ reductions today are necessary to achieve climate goals decades and centuries from now (National Research Council, 2011).

Reductions of BC today do little to achieve climate goals in the next century: however, they are important for climate goals in the near future, which can include reducing impacts on vulnerable regions such as the Arctic and reducing the rate of near-term climate change. In addition, if and when we approach climate stabilization, sustained reductions in emissions of BC will be important to keep those peak temperatures lower than they would otherwise be. Along these lines, the IPCC found that the complexity of climate change may indicate that a basket of metrics approach would best capture the variety of spatial, temporal and uncertain features (IPCC, 2009). Such a basket approach to addressing short-lived and long-lived forcers separately (though not BC specifically) has also been supported by Jackson (2009) and Daniel et al. (2011).

Outside of the policy context, the use of multiple metrics can be valuable for illustrative purposes. For example, Figure 2-26 shows the impact of BC relative to CO₂ on different timescales. Such a figure could be combined with an analysis such as the Unger et al. (2010) figure replicated in Figure 2-19 to show the GTP (or GWP) weighted impact of a set of proposed

mitigation options at 20 years and 100 years (or some other timescale).

2.7.4.1 Considering the Full Range of BC Effects

As discussed in section 2.6, BC is associated with complex indirect effects and a number of hydrological effects that are unrelated to radiative forcing and that—along with the health effects discussed in Chapter 3—distinguish it from long-lived GHGs. These effects include impacts on the water cycle, inhibition of photosynthesis due to deposition on plants (Kozlowski and Keller, 1966), enhancement of soil productivity due to deposition on soil (Laird, 2008), and effects such as surface dimming. Capturing these additional effects in a single global metric is challenging. Even the current GWP metric continues to see widespread use despite not capturing the ecosystem effects of CO₂-driven ocean acidification or the health and agricultural impacts of CH₄-induced ozone production.

For most GHGs, relative radiative forcing is a reasonable approximation of temperature impacts: a given W m⁻² of CO₂ has similar impacts to a W m⁻² of N₂O. By contrast, BC forcing includes a combination of surface dimming and absorption of both incoming and outgoing radiation at many wavelengths, while GHGs mainly absorb outgoing thermal infrared radiation. As discussed in section 2.6.1.4, the temperature change resulting from a given W m⁻² of forcing from the snow albedo effect may be much greater than the temperature change resulting from a W m⁻² of CO₂ forcing, whereas the result of forcing from BC-related direct effects may depend on the pattern of BC loading. Inclusion of the cloud effects of BC makes this metric even more uncertain.

Further complicating the use of existing metrics for BC are the significant remaining uncertainties in estimates of BC forcing, especially regarding the indirect cloud effects [which can be compared to the uncertainty in forcing from changes in well-mixed GHG concentrations, estimated to be only 10% of 2.63 W m⁻² (Forster et al., 2007)]. However, even if BC forcing is at the low end of the range, a consequence of the globally averaged nature of common metrics is that the right mix of BC and OC emissions might have little net global radiative forcing impact and yet still have significant impacts on regional precipitation, dimming, and snow melt as well as possibly on regional patterns of warming and cooling.

2.8 Key Gaps in Understanding and Expressing the Climate Impacts of BC

This chapter has summarized key findings from a wide range of peer-reviewed studies related to BC and its effects on climate. The complex atmospheric chemistry of BC and its regional nature make it a challenging subject for study. The chapter attempts to identify where the strength of the evidence suggests that reasonable conclusions can be drawn (such as for BC's direct forcing impact, which is widely understood to lead to warming), and also highlights those areas where such conclusions may be premature (such as the net effect of BC, considering its impacts on clouds and also the impact of co-emitted pollutants). Despite rapidly advancing science, there is clearly the need for additional research, particularly with regard to BC's effects on clouds and its impacts on radiative forcing, melting and precipitation in specific regions.

Recent studies have begun to apply more rigorous modeling and estimation approaches to try to provide better centralized estimates of BC's direct forcing impact, its impacts on snow and ice, and its effects on clouds. Further work is needed to improve these quantitative estimates and to ensure that the full range of BC effects on climate is considered.

Key research needs include continued investigation of basic microphysical and atmospheric processes affecting BC and other co-pollutants, particularly with regard to the climate effects of BC-cloud interactions and aerosol mixing state. In addition, there is a dearth of research on other types of light-absorbing carbon, such as BrC, which may also contribute to climate impacts especially in sensitive regions such as the Arctic. In general, further investigation of impacts of aerosols in snow- and ice-covered regions would be fruitful, along with additional research on the climate impacts of emissions mixtures from particular source categories.

It is also difficult to compare BC directly to CO₂ or other long-lived GHGs. This chapter has explored some of the metrics that are currently available to determine how well they perform for purposes of expressing the climate effects of BC and comparing BC to CO₂. However, there are clear limitations to using these metrics. In general, there is a strong need for further refinement of policy-relevant metrics for BC and other short-lived climate forcers. Appropriately tailored metrics for BC are needed in order to quantify and communicate BC's impacts and properly characterize the costs and benefits of BC mitigation.