



## Wildland fire emissions, carbon, and climate: Wildfire–climate interactions



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### ABSTRACT

Increasing wildfire activity in recent decades, partially related to extended droughts, along with concern over potential impacts of future climate change on fire activity has resulted in increased attention on fire–climate interactions. Findings from studies published in recent years have remarkably increased our understanding of fire–climate interactions and improved our capacity to delineate probable future climate change and impacts. Fires are projected to increase in many regions of the globe under a changing climate due to the greenhouse effect. Burned areas in the western US could increase by more than 50% by the middle of this century. Increased fire activity is not simply an outcome of the changing climate, but also a participant in the change. Smoke particles reduce overall solar radiation absorbed by the Earth's atmosphere during individual fire events and fire seasons, leading to regional climate effects including reduction in surface temperature, suppression of cloud and precipitation, and enhancement of climate anomalies such as droughts. Black carbon (BC) in smoke particles displays some different radiation and climate effects by warming the middle and lower atmosphere, leading to a more stable atmosphere. BC also plays a key role in the smoke–snow feedback mechanism. Fire emissions of CO<sub>2</sub>, on the other hand, are an important atmospheric CO<sub>2</sub> source and contribute substantially to the global greenhouse effect. Future studies should generate a global picture of all aspects of radiative forcing by smoke particles. Better knowledge is needed in space and time variability of smoke particles, evolution of smoke optical properties, estimation of smoke plume height and vertical profiles and their impacts on locations of warming layers, stability structure, clouds and smoke transport, quantification of BC emission factors and optical properties from different forest fuels, and BC's individual and combined roles with organic carbon. Finally, understanding the short- and long-term greenhouse effect of fire CO<sub>2</sub> emissions, increased capacity to project future fire trends (especially mega-fires), with consideration of climate–fuel–human interactions, and improved fire weather and climate prediction skills (including exploring the SST–fire relations) remain central knowledge needs.

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### 1. Introduction

Wildfires and climate are two closely related Earth system processes. It has long been recognized that atmospheric conditions are an environmental factor for wildfires. Depending on their time scales, atmospheric conditions are classified into weather and climate. Weather is commonly defined as the day-to-day state (temperature, humidity, wind, etc.) and processes (cloud and precipitation, fronts, jets, troughs, ridges, etc.) of the atmosphere in a region and their short-term (up to weeks) variations, whereas climate is defined as statistical weather information over a certain period (usually 30 years) ([www.diffen.com/difference/Climate\\_vs\\_Weather](http://www.diffen.com/difference/Climate_vs_Weather)). Climate also generally serves as a reference to atmospheric variability on time-scales that exceed the limit of

deterministic predictability, about 2–3 weeks (Wallace and Hobbs, 2006).

Atmospheric conditions for fires are accordingly classified into fire weather and fire climate (Pyne et al., 1996). Fire weather is one of the factors determining occurrence and behavior of individual fires within a fire season (Flannigan and Wotton, 2001). Fire climate, meanwhile, is a synthesis of daily fire weather that describes statistical features (average, variation, etc.) of fire weather. Fire climate determines the atmospheric conditions for fire activity at time scales beyond a fire season (Flannigan and Wotton, 2001).

Wildfires can impact atmospheric conditions at various spatial and temporal scales through emissions of gases, particles, water, and heat. Fire emission components (Table 1) with significant atmospheric effects include CO<sub>2</sub> (about 71% of mass), and organic and element or black carbon accounting for about 0.24% and 0.02%, respectively. Organic carbon and black carbon are carbonaceous aerosols that scatter and absorb atmospheric radiation, respectively. The percentage of a component varies considerably

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**Table 1**

Emissions from wildfires. The mass amount is grams of emission/kilogram of fuel burned. The letter “d” represents particle diameter (from NRC, 2004).

| Emissions                               | Mass   | %     |
|---|--------|-------|
| Carbon Dioxide                          | 1564.8 | 71.44 |
| Carbon Monoxide                         | 120.9  | 5.52  |
| Organic Carbon                          | 5.2    | 0.24  |
| Elemental Carbon                        | 0.4    | 0.02  |
| Particulate matter $d < 2.5\mu$         | 10.3   | 0.47  |
| Particulate matter $2.5\mu < d < 10\mu$ | 1.9    | 0.09  |
| Particulate Matter $d > 10\mu$          | 3.8    | 0.17  |
| Nitric Oxide                            | 8.5    | 0.39  |
| Methane                                 | 5.9    | 0.27  |
| Non-methane Hydrocarbon                 | 4.3    | 0.20  |
| Volatile Organic Compounds              | 5.2    | 0.24  |
| Water                                   | 459.2  | 20.97 |

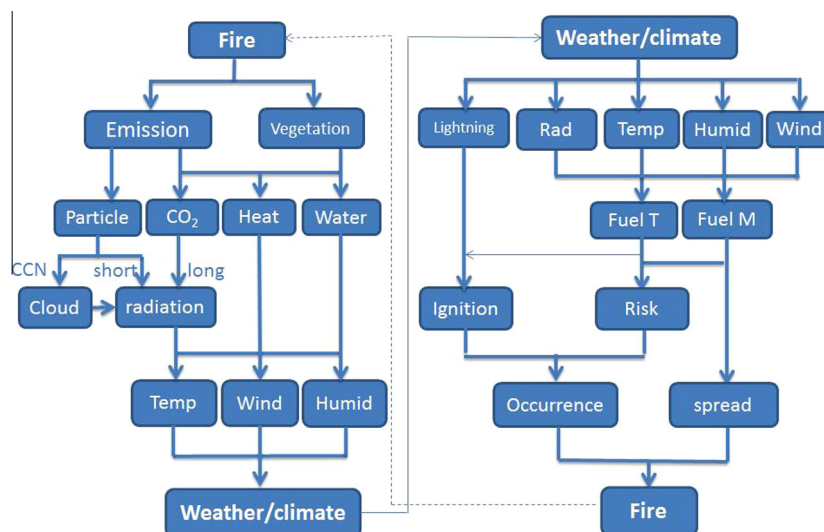
from one burn case to another. For example, CO<sub>2</sub> reported by Urbanski et al. (2008) accounts for 87–92% of total carbon burned. Smoke particles can directly affect atmospheric radiative transfer through scattering and absorbing radiation (mainly short-wave radiation) and indirectly through changing cloud properties with smoke particles acting as cloud condensation nuclei (CCN) (Fig. 1). Changes in radiative forcing lead to subsequent changes in air temperature, humidity, and wind. The changes can happen at short time scales of minutes and days. Additionally, CO<sub>2</sub> is a dominant component of fire emissions. As a greenhouse gas (GHG), CO<sub>2</sub> absorbs atmospheric long-wave radiation emitted from the surface-atmosphere system and therefore is a primary factor in global warming. This usually happens over a long-term period of decades. Heat energy released from fires can also modify the local atmospheric thermodynamic structure, turbulence regime, and wind patterns, as well as other atmospheric thermal and dynamical properties and processes. Water vapor released from fire can increase atmospheric humidity, favoring formation of clouds and fog. Similar to smoke particles, the released heat and water can affect atmospheric conditions at short-time scales. Fires also remove certain amount of vegetation coverage or reduce vegetation density, which will affect heat, water, and trace gas exchanges with the atmosphere at both short- and long-term scales, which will in turn affect weather and climate.

Gas and particle emissions from fires alter atmospheric properties (air temperature, humidity, clouds, wind and turbulence) which in turn will modify fuel conditions, especially fuel moisture,

at different time and space scales by. For example, heat release has an immediate impact on winds which changes the surface fluxes of heat and moisture and thereby alters fuel moisture and fuel temperature, which are important factors for fire occurrence (ignition and risk) and fire spread. Solar heating of fuel particles is a prime driver of fuel temperature and fuel moisture (see e.g. Rothermel et al., 1986; Cohen and Deeming, 1985; Carlson et al., 2007). Changes in atmospheric moisture content and cloud properties due to fire emissions can alter the amount of solar radiation reaching the surface. Fuel particles absorb solar radiation resulting in a temperature increase of the fuels. Also, exposed ground surfaces are also heated by solar radiation and subsequently transfer heat to adjacent fuel particles. However, CO<sub>2</sub> emissions contribute to global CO<sub>2</sub> levels and impact climate over a decadal or longer time-scale. Future changes in climate may impact future fire behavior. Smoke particles can have immediate (shading of fuels from solar radiation, impact on local atmospheric structure) and longer-term impacts that act over larger spatial scales on fire behavior (aerosol impacts on radiation and cloud processes may act over time periods of hours to days via impact larger areas or be realized some distance from the fire source).

While research has historically focused on the fire–weather interactions, increasing attention has been paid in the past few decades to fire–climate interactions. A contributing factor to this emerging emphasis is the evidence that wildfires, especially large wildfires, have increased in recent decades (Piñol et al., 1998; Goldammer, 2001; Gillett et al., 2004; Reinhard et al., 2005; Westerling et al., 2006), partially related to extreme weather events such as extended droughts (Goldammer and Price, 1998; Stocks et al., 2002). Persistent weather anomalies can directly impact fire activities during a fire season. Under prolonged warm and dry conditions, fires are easier to ignite and spread and a fire season often becomes longer. The duration of the fire season may also be impacted by earlier beginning due to low snow pack or early spring snow melt (Westerling et al., 2006). Another factor is concern regarding the potential impacts of future climate change on fire activity. Many climate models have projected significant climate change during this century due to the greenhouse effect (IPCC, 2007), including an overall increase in temperature worldwide and a drying trend in many subtropical and mid-latitude regions. It appears likely that wildfires will increase in these regions.

Knowledge of fire–climate interactions is essential to understanding fire and climate variability and change. First, seasonal



**Fig. 1.** Diagram of physical processes for fire's impacts on weather and climate and feedbacks.

predictions of fire risk provide invaluable aid to fire and land managers in planning fire suppression and other fire-related activities (Westerling et al., 2002; Brown et al., 2004; Roads et al., 2005). Secondly, understanding the mechanisms for seasonal atmospheric anomalies such as droughts is an extremely important climate issue with relevance beyond the fire community. Such anomalies are largely driven by interactions within the climate system such as air–sea and air–land interactions (e.g., Trenberth et al., 1988; Giorgi et al., 1996). Fire–climate interactions suggest that fire emissions are a possible external factor, which, when understood, would contribute to improving prediction skills. In addition, projection of future wildfire trends under a changing climate is essential to assessing potential impacts of wildfires on human social systems and the environment and is critical to designing and implementing necessary measures to mitigate these impacts.

This paper synthesizes studies on fire–climate interactions. The issues to be addressed include radiative forcing of fire emissions, climatic impacts, fire prediction, and future trends of fires under a changing climate. These issues have been reviewed or synthesized in many studies, including Flannigan and Wotton (2001) for fire–weather/climate interactions in Canada and US; Kanakidou et al. (2005) and Ramanathan and Carmichael (2008) for radiative and climatic impacts of aerosol (organic aerosol and black carbon), respectively; Flannigan et al. (2009a) for fire and climate change; Bowman et al. (2009) for the role of fire in the Earth system; Langmann et al. (2009) for fire emissions and climatic and air quality impacts; and Hessl (2011) for climate and fire regimes. This synthesis focuses on radiative forcing and climatic impacts of smoke emissions, future fire trends under the changing climate, and the studies for the United States.

## 2. Radiative forcing of fire emissions

### 2.1. Fire emissions

Carbon is one of the important fire emission components and is mainly included in both gases (e.g., CO<sub>2</sub>, CO, and CH<sub>4</sub>) and particles (organic carbon and black carbon). Global carbon emissions have been estimated during the past three decades (e.g., Seiler and Crutzen, 1980; Crutzen and Andreae, 1990; Dixon and Krankina, 1993; Hao et al., 1996; Galanter et al., 2000; Andreae and Merlet, 2001; Amiro et al., 2001; Page et al., 2002; Schultz, 2002; Duncan et al., 2003; Mouillot et al., 2006; Schultz et al., 2008; van der Werf et al., 2010; Mieville et al., 2010; Wiedinmyer et al., 2011). The first complete estimate of averaged annual global carbon emissions was 2.6 Pg C yr<sup>-1</sup> (Seiler and Crutzen, 1980). Most recent estimates ranged from about 1.4 Pg C yr<sup>-1</sup> (Ito and Penner, 2004) to 2.8 Pg C yr<sup>-1</sup> (Langmann et al., 2009). In the latest Global Fire Emission Dataset (GFED3, van der Werf et al., 2010), an estimate of 2.0 Pg C yr<sup>-1</sup> was obtained for the period of 1997–2009 using a biogeochemical model and satellite estimates of burned area (Giglio et al., 2010) and productivity, which is about one-third of the total carbon emissions. This contribution could be extremely significant over a short period of time before carbon uptake resulting from regrowth of burned area vegetation. The estimated carbon emissions during the 1997–98 Indonesian wildfires were the equivalent to the total global carbon uptake by the terrestrial biosphere in a typical year (Page et al., 2002; Tacconi et al., 2006). The contribution could be significant also over a longer period because a large portion of carbon stored in forest and other ecosystems could be lost permanently in many regions such as the Amazon where deforestation was achieved using biomass burning.

Black carbon is a product of incomplete combustion, which together with organic carbon constitutes the majority of particulate carbon. Approximately 5–10% of fire smoke particles are BC, as

compared to about 50–60% for OC, and the variations can be due to different fuel types and combustion conditions, as well as the analytical methods used (Reid et al., 2005a). The OC/BC ratios in Wiedinmyer et al. (2011) were about 10:1, and their total amounts were about half of those of PM<sub>10</sub>. According to IPCC (2007), global fossil fuel emission estimates of BC at present range from 5.8 to 8.0 Tg C yr<sup>-1</sup> (Haywood and Boucher, 2000). Biomass burning (forest and savanna burning) contributes about 40% of total BC emissions. Bond et al. (2004) estimated the total current global emission of BC to be 4.6 Tg C yr<sup>-1</sup> from fossil fuel and biofuel combustion and 3.3 Tg C yr<sup>-1</sup> from open biomass burning. Wiedinmyer et al. (2011) estimated the global fire BC emission of 2.2 Tg C yr<sup>-1</sup> for the period of 2005–2010, which is comparable to that from The Global Fire Emissions Database (GFED) (Randerson et al., 2005; van der Werf et al., 2004, 2006, 2010).

One of the important properties of smoke as well as other tropospheric aerosols is its large spatial variability due to the local and regional origination of fires, short periods of individual fire events, and short lifetime of particles after being emitted into the atmosphere. Measurements have shown dramatically large amounts of wildland fire emission in the Amazon and North America (e.g., Radke, 1991; Ward and Hardy, 1991; Liu, 2004) during individual fires or a burning season, making smoke an important factor to radiation budget in these regions. Biomass burning in the tropics is of particular interest because of the large extent of forest clearing and agricultural burning.

Carbon emission from fires in the major geographical regions or ecoregions in North America have been estimated in many studies (e.g., French et al., 2004, 2011; Amiro et al., 2001; Kasischke and Bruhwiler, 2002; Ito and Penner, 2004; Hoelzemann et al., 2004; Liu, 2004; Liu et al., 2005a; Kasischke and Johnstone, 2005; Wiedinmyer et al., 2006; Schultz et al., 2008; Reid et al., 2009; van der Werf et al., 2010). The estimates from van der Werf et al. (2010) are on the order of 10 Tg C yr<sup>-1</sup> for the continental US and Mexico and 50 Tg C yr<sup>-1</sup> for Canada and Alaska.

### 2.2. Impacts of smoke emissions on atmospheric radiation

Incident solar radiation that drives the earth's climate system is either reflected back to space (~30%) or absorbed by the earth's surface and atmosphere (~70%). It is this absorbed radiation that heats the planet and atmosphere (Ramanathan and Feng, 2009). The overall energy budget for the planet includes not only the amount of solar radiation absorbed and reflected by the earth's surface and atmosphere, but also the amount of absorbed radiation re-emitted from the earth's surface and atmosphere as long wave radiation. Greenhouse gases and aerosols produced from wildland fires and generated in the atmosphere through chemical reactions involving precursor chemicals emitted from those fires affect the earth's overall energy balance (and thus temperature) because they also absorb and reflect long-wave and solar radiation, clouds with the roles of smoke particles as CCNs. The changes in radiation and clouds can further affect energy and water conditions for the soil-vegetation system.

As one of the sources of atmospheric aerosols (Andreae and Merlet, 2001), smoke particles can produce radiative forcing through three mechanisms. First, as with other types of atmospheric aerosols, smoke particles can impact shortwave radiation as well as long-wave radiation through scattering and absorbing solar radiation, a mechanism known as “direct radiative forcing” (DRF) (Charlson et al., 1992), or “aerosol direct effect” (IPCC, 2007). Second, clouds are an important factor for atmospheric radiation transfer. Serving as CCNs, aerosols including smoke particles modify the microphysical and hence the radiative properties, amount and lifetime of clouds, a mechanism called “indirect radiative forcing” (IRF) (e.g., Twomey et al., 1984; Kaufman and Tanré,

1994), or “aerosol indirect effect” (IPCC, 2007). The microphysically induced effect on the cloud droplet number concentration and hence the cloud droplet size, with the liquid water content held fixed has been called the “cloud albedo effect”, while the microphysically induced effect on the liquid water content, cloud height, and lifetime of clouds has been called specific names such as the “cloud lifetime effect” (IPCC, 2007). Third, the aerosol radiative forcing (both DRF and IRF) will change atmospheric structure, circulation, and energy and water exchanges on the ground surface. This will affect atmospheric water vapor and clouds and will further affect radiation, a mechanism known as “semi-direct radiative forcing” (Hansen et al., 1997), or “aerosol semi-direct effect” (IPCC, 2007).

### 2.3. Optical properties of smoke particles

The impacts of atmospheric aerosols including smoke particles on radiation depend on their optical properties, mainly aerosol optical depth (AOD) and single scattering albedo (SSA) (Hansen et al., 1997; Kanakidou et al., 2005). The optical properties of aerosol particles depend strongly on the size distribution, morphology, chemical composition, and mixing states (Reid et al., 2005a,b; Jacobson, 2001). AOD is the extinction resulting from absorption and scattering of radiation by the aerosol in a column. SSA is the ratio of scattering to the sum of scattering and absorption and is an indicator of intensity of absorption capacity of aerosol. A value of unity (one) represents pure scattering aerosols. The smaller the SSA value is, the stronger the aerosol absorption is. Aerosol optical properties depend on particle size distribution and radiation wavelength.

The magnitude and variations of these two optical properties have been reported for wildfires and controlled biomass burning in a number of climate zones. The AOD of 0.75 (Ross et al., 1998) and SSA from 0.82 for young smoke and 0.94 for aged smoke (Eck et al., 1998) at about 550 nm were obtained from the Smoke, Clouds and Radiation-Brazil (SCAR-B) during the 1995 biomass burning season in Amazon. Comparable values were also obtained from the Aerosol Robotic Network (AERONET) measurements during the SAFARI 2000 dry season campaign in southern Africa (Eck et al., 2003). The maximum AOD at 550 nm ranged from 0.52 to 0.87 with SSA at 440 nm ranging from 0.92 to 0.98 for observations at a number of sites across eastern Europe, northern Scandinavia, and Svalbard near the Arctic (Lund Myhre et al., 2007). The mean AOD at 500 nm for April 2008 at two AERONET boreal sites in Alaska was 0.28 with maximum daily values of about 0.8 and SSA at 440 nm ranged from 0.91 to 0.99 with an average of 0.96 for observations in 2004 and 2005 (Eck et al., 2009).

The optical properties of smoke may differ substantially among different climate zones and smoke age. Eck et al. (2003) compared optical properties of four biomass burning events (Table 2). Two of them (Zambia and Brazil) were tropical fires with the fuel types of savanna and mixed forest and pasture. Others (Maryland and Moldova) were boreal fires with the fuel types of forest and peat. There are noticeable differences between the tropical and boreal fires.

The particle sizes with the largest volume are smaller for the tropical fires (0.15 and 0.18  $\mu\text{m}$ ) than the boreal fires (0.2 and 0.25  $\mu\text{m}$ ). The smoke was a mixture of young and aged for the tropical fires, but aged for the boreal ones. AOD, which decreases with wavelength, is the same for all fire cases at 500 nm, but larger (smaller) at the wavelengths  $>500$  nm ( $<500$  nm) for the tropical fires than the boreal ones. SSA, which is much less dependent on wavelength, is smaller for the tropical fires than the boreal ones, indicating that young smoke has stronger absorption than aged smoke.

AOD of smoke particles increases with humidity (e.g., Jeong et al., 2007), a hygroscopic property. As relative humidity (RH) increases, aerosols absorb water from the air, which increases the particle size and therefore increases the particle scattering cross section. Hygroscopic growth is only important for RH greater than about 40% and, at a given RH, varies with particle solubility (Reid et al., 2005b).

### 2.4. Radiative forcing of smoke particles

A metric typically used to assess and compare the anthropogenic and natural drivers of climate change, including greenhouse gases, aerosols, and black carbon, is radiative forcing (Forster et al., 2007). The definition of radiative forcing as adopted by the Intergovernmental Panel on Climate Change (IPCC) is the change in net radiation ( $\text{W m}^{-2}$ ) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium (Ramaswamy et al., 2001; Forster et al., 2007).

The IPCC reports provided estimates of direct radiative forcing associated with the emissions of principal gases and aerosols (including aerosol-precursors). The emissions of aerosols generally contribute to a negative radiative forcing through the scattering of solar radiation. In the third assessment report (TAR) (IPCC, 2001a), DRF was estimated to be  $-0.4$ ,  $-0.1$ , and  $+0.2 \text{ W m}^{-2}$  from sulfate, fossil OC, and fossil BC aerosols, respectively, emitted during the period of 1750–1998 (Table 3). The TAR reported a contribution of biomass burning to the DRF of roughly  $-0.4 \text{ W m}^{-2}$  from the scattering components (mainly organic carbon and inorganic compounds) and  $+0.2 \text{ W m}^{-2}$  from the absorbing components (BC), leading to an estimate of the net DRF of biomass burning aerosols of  $-0.20 \text{ W m}^{-2}$ . In the IPCC fourth assessment report (FAR) for the aerosols emitted during the period of 1750–2005 (Forster et al., 2007; IPCC, 2007), DRF remained the same for sulfate and fossil BC aerosols, but the magnitude was slightly reduced to  $-0.05 \text{ W m}^{-2}$  for fossil OC. The FAR estimate of the net DRF from biomass burning aerosols turned to slightly positive at  $0.03 \text{ W m}^{-2}$ . The change was mainly owing to improvements in the models in representing the absorption properties of the aerosol and the effects of biomass burning aerosol overlying clouds.

The large amount of incident solar radiation in the tropics enhances the radiative forcing of aerosols (Holben et al., 2001). Penner et al. (1992) emphasized the importance of smoke particles in the Amazon to the global radiative budget. Based on carbon emissions from biomass burning (Crutzen and Andreae, 1990;

**Table 2**

Comparisons of smoke aerosol optical depth (AOD) and single scattering albedo (SSA) between tropical and boreal wildfires with different smoke ages. “F & A” represents fresh and aged. “T” and “B” represent tropical and boreal, respectively. The numbers right below AOD and SSA are wavelengths (the data was from Eck et al., 2003).

| Location, date   | Vegetation type     | Smoke age | Size ( $\mu$ ) | AOD |     |     |     | SSA   |       |       |       |
|------------------|---------------------|-----------|----------------|-----|-----|-----|-----|-------|-------|-------|-------|
|                  |                     |           |                | 440 | 500 | 675 | 870 | 440   | 500   | 670   | 870   |
| Zambia, 9/15/00  | T. savanna          | F & A     | 0.15           | 2.5 | 2.1 | 1.2 | 0.7 | 0.88  | 0.87  | 0.85  | 0.82  |
| Brazil, 8/15/02  | T. forest & pasture | F & A     | 0.18           | 2.5 | 2.1 | 1.1 | 0.6 | 0.935 | 0.93  | 0.925 | 0.92  |
| Maryland, 7/8/02 | B. forest           | 2.5 days  | 0.2            | 2.3 | 2.1 | 1.4 | 0.9 | 0.96  | 0.965 | 0.97  | 0.965 |
| Moldova, 9/11/02 | B. forest & peat    | >2 days   | 0.25           | 2.2 | 2.1 | 1.5 | 1.0 | 0.99  | 0.99  | 0.99  | 0.99  |

**Table 3**

Direct radiation forcing (or aerosol direct effect) in  $\text{W m}^{-2}$  from various sources in the IPCC Third Assessment Report (TAR, IPCC, 2001a) and Fourth Assessment Report (FAR, IPCC, 2007).

| Source  | TAR  | FAR   |
|---------|------|-------|
| Sulfate | -0.4 | -0.4  |
| Fossil  |      |       |
| OC      | -0.1 | -0.05 |
| BC      | +0.2 | +0.2  |
| Biomass |      |       |
| OC      | -0.4 | +0.03 |
| BC      | +0.2 |       |

Hao et al., 1990), a globally averaged smoke DRF of about  $-1 \text{ W m}^{-2}$  was obtained, comparable to that of anthropogenic sulfate aerosols. Hobbs et al. (1997) reassessed the role of smoke from biomass burning using airborne measurements in Brazil and obtained a value that is only about one-third of the early estimate. However, they pointed out that the DRF could be larger on regional scales. This result was also confirmed in Ross et al. (1998), who obtained DRF of  $-15 \pm 5 \text{ W m}^{-2}$  for the 1995 Amazon smoke season using a one-dimensional atmospheric radiative transfer model with a total optical depth of 0.75. This magnitude is equivalent to an annually averaged DRF of about  $-2.5 \text{ W m}^{-2}$  in a typical smoke area in Brazil. Large smoke DRF was also found in Africa during the Southern African Regional Science Initiative (SAFARI 2000) (Swap et al., 2003), in Southeast Asia during the 1997 forest fires (Kobayashi et al., 2004), and in the 1988 Yellowstone fires (Liu, 2005a).

The magnitude of IRF may be comparable to or even greater than that of DRF. In the IPCC FAR (IPCC 2007), the IRF of all atmospheric aerosols emitted during the period of 1750–2005 was estimated to be  $-0.7 \text{ W m}^{-2}$  with a range from  $-0.18$  to  $-0.9 \text{ W m}^{-2}$ . Chuang et al. (2002) estimate an indirect aerosol forcing of  $-1.16 \text{ W m}^{-2}$  for carbonaceous aerosols from fires, although this estimate only includes the cloud albedo effect. Ward et al. (2012) included additional indirect effects such as effects on cloud height and lifetime, and showed comparable forcings, ranging between  $-1.74$  to  $-1.00 \text{ W m}^{-2}$ .

There are no estimates yet for the semi-direct radiative forcing in the IPCC reports. However, a few case studies provided some estimates of its magnitude. Liu (2005b) obtained a DRF of  $-16.5 \text{ W m}^{-2}$  for the smoke particles from the Amazon biomass burning simulated with a three-dimensional regional climate model. The magnitude is sharply reduced to  $-9.8 \text{ W m}^{-2}$  over the smoke region when the atmospheric feedback of reduced clouds is considered. The semi-direct radiative forcing is therefore about  $+7 \text{ W m}^{-2}$ .

### 2.5. Black carbon

A special optical property of BC that differentiates it from other types of carbonaceous aerosol is its strong absorption of solar radiation. Thus, although the overall radiative forcing of atmospheric aerosol is negative, the BC component can produce positive radiative forcing. (IPCC, 2007). Chung et al. (2005) and Ramanathan and Carmichael (2008) reported a global total black carbon DRF of  $0.9 \text{ W m}^{-2}$ , a value larger than the IPCC estimates and the DRF associated with other greenhouse gases such as  $\text{CH}_4$ ,  $\text{N}_2\text{O}$ , or tropospheric  $\text{O}_3$ . A number of other studies (e.g., Haywood and Ramanaswamy, 1998; Jacobson, 2001; Chung and Seinfeld, 2005; Sato et al., 2003; Bond et al., 2010) also reported large radiative forcing values between  $0.4$  and  $1.2 \text{ W m}^{-2}$  from all BC emissions. Chung and Seinfeld (2005) estimated a radiative forcing range of  $0.52$ – $0.93 \text{ W m}^{-2}$  for the Northern Hemisphere due to black carbon

emissions from fossil fuels, biofuels, and biomass burning. Myhre et al. (2009) reported a radiative forcing range of about  $0.1$ – $0.7 \text{ W m}^{-2}$  over the contiguous US due to fossil fuel and biofuel emissions. For BC emitted from biomass burning alone, the global radiative forcing was estimated to be  $+0.2 \text{ W m}^{-2}$  in the IPCC TAR (IPCC, 2001b).

### 2.6. $\text{CO}_2$ radiative forcing

According to measurements recorded at a Hawaiian observatory, atmospheric  $\text{CO}_2$  concentrations rose from  $315.98 \text{ ppmv}$  in 1959 to  $385.34 \text{ ppmv}$  in 2008 (Keeling et al., 2009), a 22% increase over 50 years. The concentrations have increased by about 40% from about  $285 \text{ ppmv}$  in the mid-1700s. Atmospheric  $\text{CO}_2$  can absorb long-wave radiation emitted from the ground. The IPCC FAR (IPCC 2007) estimated that the radiative forcing resulting from  $\text{CO}_2$  increases since 1750 is about  $1.66 \pm 0.17 \text{ W m}^{-2}$ . The large ratio of fire to total carbon emissions suggests a significant contribution of fire to total  $\text{CO}_2$  radiative forcing.

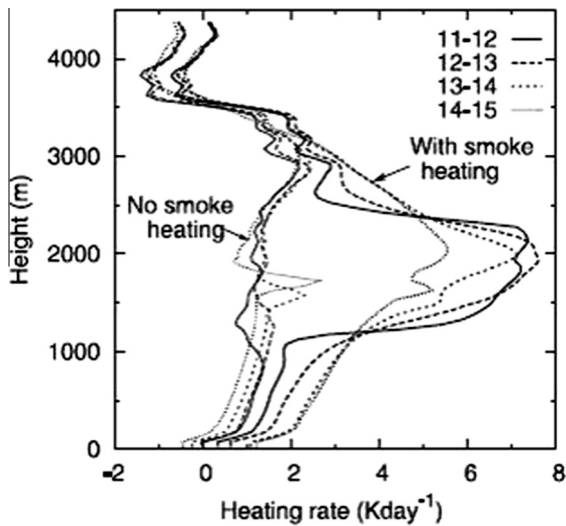
## 3. Climatic impacts

### 3.1. Smoke particles

#### 3.1.1. Atmospheric thermal structure and circulations

Solar radiation is the ultimate energy source for the atmosphere and one of the energy balance components that determine atmospheric thermal structure. While the impact of GHG (including those from fire emissions) increases on atmospheric warming has been effectively determined for some time, the impact of aerosols (including those from fire emissions) has been more difficult to quantify. Aerosol radiative forcing impacts, direct and indirect, demonstrate significant variability in space and time, with current estimates indicating negative forcing for both aerosol direct and indirect (cloud albedo) forcing (Quaas et al., 2008). Estimates are that anthropogenic aerosols have overall lessened the warming impacts resulting from GHG increases and have had a significant impact on critical ocean–atmosphere interactions that drive important cycles of atmospheric variability (Evan et al., 2011; Booth et al., 2012; Evan, 2012). Radiative forcing of smoke particles is negative when smoke particles are locally present in sufficient density in the atmosphere, meaning that as radiation absorbed by the earth–atmosphere system becomes smaller, and the ground surface will experience cooling. This was observed during a wild-fire near Boulder, Colorado, in 2010 (Stone et al., 2011) where the surface under the smoke plume was cooled by  $2$ – $5 \text{ }^\circ\text{C}$ . Note that an individual fire event like the Colorado fire may last only for a few days, meaning that it would mostly affect weather processes. However, there are usually many such individual fires in a specific region during a fire season, which together can affect regional climate.

In another fire case, smoke aerosol in a daytime convective boundary layer was found to warm the atmosphere (Fig. 2) and affect cloud formation, and the vertical distribution of smoke aerosol in the convective boundary layer was found to be crucial to determining whether cloudiness is reduced (Feingold et al., 2005). The warming due to solar radiation absorption in the atmospheric smoke, coupled with the cooling on the land surface and the atmosphere below the smoke layer due to solar radiation scattering and absorption by smoke, would make the atmosphere more stable and, therefore, suppress cloud development. The net change in air temperature depends on the relative importance of the absorption and the change in sensible heat flux on the ground surface related to the reduction in solar radiation absorbed due to smoke particles (Liu, 2005b). The lower and middle troposphere



**Fig. 2.** Vertical profiles of radiative heating rate in Amazonia simulated with an atmosphere-cloud model. Hourly averages were made from 1100 through 1500 local time for simulation with and without smoke, respectively (from Feingold et al., 2005).

becomes more thermally stable. The net cooling effect of smoke particles may have implications for climate change. Smoke transported from wildfires in northern boreal forests to the Arctic could cool the Arctic for weeks to months at a time, temporarily counteracting warming due to the greenhouse effect (Stone et al., 2008).

The change in atmospheric thermal structure due to the direct radiative forcing of smoke particles can further change regional circulations (Liu et al., 2005a; Liu, 2005b; Evan et al., 2011; Booth et al., 2012). Simulations (Liu et al., 2005a; Liu, 2005b) showed that smoke particles emitted during a biomass burning season in South America increased 500 hPa geopotential heights over the smoke region, indicating a tendency of enhanced Atlantic Ocean high or weakened tropical trough. This tendency could last into post-burn period, implying a delay of monsoon onset or weakening of its intensity.

### 3.1.2. Clouds and precipitation

Clouds and precipitation are usually reduced in the presence of smoke particles (Ackerman et al., 2000; Koren et al., 2004; Andreae et al., 2004; Liu, 2005b). Different physical mechanisms have been proposed for this reduction. During the Indian Ocean Experiment (INDOEX) there was relatively small cloud coverage over the ocean area due to the large concentration of soot (a substance consisting of pure light-absorptive BC and partial light-absorptive carbon or brown carbon) aerosols, which increase air temperature, reduce relative humidity, and therefore “burn out” clouds (Ackerman et al., 2000). Cloud and precipitation reductions due to smoke over the Amazon were found mainly as a result of smaller water vapor transport from the ground and the planetary boundary layer to the cloud layer due to the combined effects of reduced turbulent activity and the subsidence tendency (Liu, 2005b). For the intense wildfires during the 2004 Alaska fire season, the high concentrations of fine aerosol ( $PM_{2.5}$ ) and the resulting large numbers of CCN had a strong impact on cloud microphysics when clouds were present, with decreased or increased precipitation, depending on the time into model simulation employed (Grell et al., 2011). The cloud impact of smoke also depends on intensity of smoke radiative forcing (Ten Hoeve et al., 2012). With increasing AOD, cloud optical depth (COD) was found to decrease at higher AODs, but increase at lower AODs. Field measurements provided observational evidence for cloud changes directly related to biomass burning in

the Amazon region (Koren et al., 2004; Andreae et al., 2004). Koren et al. (2004) analyzed the satellite measurements from Moderate Resolution Imaging Spectroradiometer (MODIS) during the biomass burning season and found that cloud coverage was reduced from 38% in clean conditions to almost 0% for heavy smoke (Fig. 3).

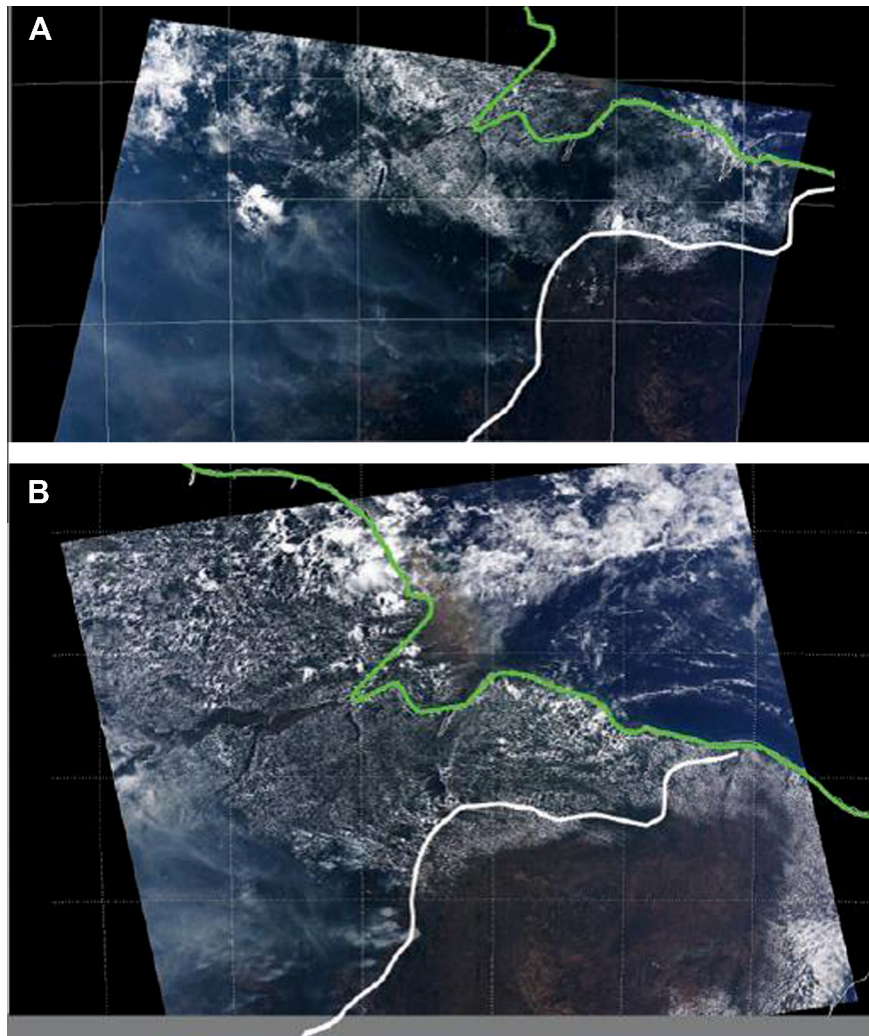
### 3.1.3. Fire weather

As the previous sections have pointed out, wildland fires are capable of altering local weather and climate conditions. However, these changed atmospheric conditions also feedback upon fires. Werth et al. (2011) provides a thorough review of critical fire weather patterns associated with extreme fire behavior. These critical patterns act to bring into alignment the important weather elements for destructive wildfires: hot, dry, and windy conditions. The role of particulate emissions from fires influencing cloud properties and altering precipitation patterns provides the most direct link to hot and dry conditions through reduced cloud cover and rainfall amounts. The impact of changes in either surface temperature or the vertical profile of temperature is not as simple.

Changes in surface air temperature, humidity and wind are not a primary driver of fire behavior on their own, but rather act through changes to fuel moisture. Assuming the moisture content of the air remains constant, an increase in temperature results in decreases in fuel moisture and therefore enhanced fire behavior. Shading from a smoke plume can reduce surface temperatures, thereby increasing fuel moisture levels and suppressing the fire behavior. Other areas not directly shaded by the smoke column often experience reduced cloud cover and therefore receive more incoming solar radiation at the surface, which increases the surface temperature. The greenhouse gas component of wildfire smoke's impact on the climate system suggests that surface temperatures would be warmer, leading one to expect reduced fuel moistures and enhanced fire behavior. However, warmer global temperatures also imply increased evaporation from the oceans adding moisture to the lower levels of the atmosphere and perhaps increasing cloud cover, which makes estimating the impact of greenhouse gases induced warming on fire behavior difficult. Warming at upper levels of the troposphere due to the absorption of solar radiation by smoke particles will act to make the atmosphere more stable. While stable atmospheric conditions are not favorable for extreme fire behavior in most cases, these conditions are favorable for the cloud free conditions that generally support the requirements of a hot and dry environment.

### 3.1.4. Seasonal climate anomalies

For fire seasons that often are associated with droughts, the impacts of smoke particles on radiation and precipitation can last for several months and therefore may reinforce seasonal climate anomalies. Liu (2005a) simulated the role of the Yellowstone fires in the development of the 1988 northern US drought using a regional climate model. The precipitation perturbation in response to radiative forcing of smoke aerosols is mostly negative in the Northwest, with the largest reduction of about  $-30$  mm in the Great Lakes region. The simulated perturbation pattern is similar to the observed pattern of precipitation anomalies, suggesting that the smoke particles might have enhanced the drought (Fig. 4). Tosca et al. (2010) investigated the interactions between equatorial Asian fires and ENSO-induced regional drought using satellite observations and atmospheric modeling of several types of smoke affected radiative forcing and precipitation variations, and found that the combination of decreased SSTs and increased atmospheric heating reduced regional precipitation. The vulnerability of ecosystems to fire was enhanced because the decreases in precipitation exceeded those for evapotranspiration. The results imply a possible positive feedback loop in which anthropogenic burning intensified drought stress regionally during El Niño.



**Fig. 3.** MODIS images of the east Amazon basin on August 11, 2002, showing beginning of cloud formation in the morning (A) and full development of clouds in the afternoon except for the smoke areas (the portion below the white line) (from Koren et al., 2004).

### 3.2. Black carbon

BC emissions enhance the greenhouse effect in the atmosphere, which is mainly caused by the increased atmospheric  $\text{CO}_2$  concentration. Due to the strong solar radiation absorption capacity and high concentrations at tropical latitudes where solar irradiance is highest, black carbon emissions are considered to be the second strongest contributor to current global warming, after  $\text{CO}_2$  emissions (Ramanathan and Carmichael, 2008). According to their estimate, the radiative forcing of total BC emissions would have a globally averaged surface warming effect of 0.5–1.0 °C. The role of BC in climate change was emphasized in the recent EPA's report on BC to Congress (EPA, 2012).

A special role of BC in climate variability and change is related to BC-snow interactions. The deposition of BC transported from other parts of the world on snow and ice at high latitudes reduces albedo and increases solar radiation absorbed by the surface, which in turn accelerates snow melting (Hansen and Nazarenko, 2004). Boreal fires contribute more BC to the Arctic than anthropogenic sources in summer based on multiyear averages (Stohl et al., 2006). A case study of extensive boreal fires in Russia during 2003 estimated that they contributed 50% of the total BC deposited north of 75 N in spring and summer and were a big factor for local haze (Fig. 5) (Generoso et al., 2007). Flanner et al. (2007) indicated

that global land and sea-ice snowpack absorbed 0.60 and  $0.23 \text{ W m}^{-2}$ , respectively, because of direct BC/snow forcing in a strong fire year.

### 3.3. Greenhouse effects of $\text{CO}_2$ emissions

Greenhouse effects and related climate change with the increased atmospheric  $\text{CO}_2$  concentrations have been extensively assessed in the IPCC reports (IPCC, 2001b, 2007), which basically indicate warming worldwide, overall drying in many subtropical and mid-latitude regions, and more frequent and intense climate anomalies. This provides some essential information for assessing the climatic effects of fire emissions, which are one of the important sources for atmospheric  $\text{CO}_2$ .

### 3.4. Fire-atmosphere interaction through land cover change

Wildfire is a disturbance of ecosystems. Thus, besides the atmospheric impacts through particle and gas emissions, wildfires also affect the weather and climate through modifying terrestrial ecosystem structure, processes, and services such as carbon sequestration, soil fertility, grazing value, biodiversity, and tourism, and can hence trigger land use change (Lavorel et al., 2007). The land surface is the underlying boundary of the atmosphere and can af-

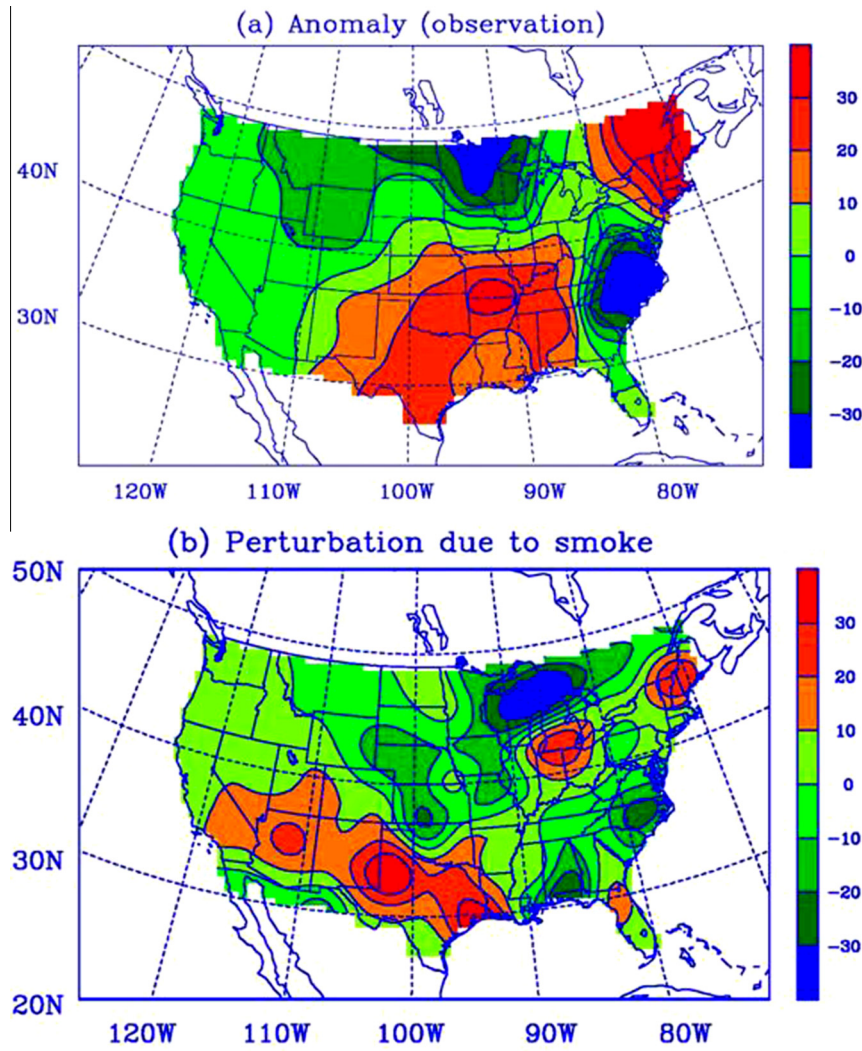


Fig. 4. Precipitation anomalies (mm) in July 1988. (a) Observation, and (b) Difference between the regional climate model simulations with and without smoke particles (from Liu, 2005a,b).

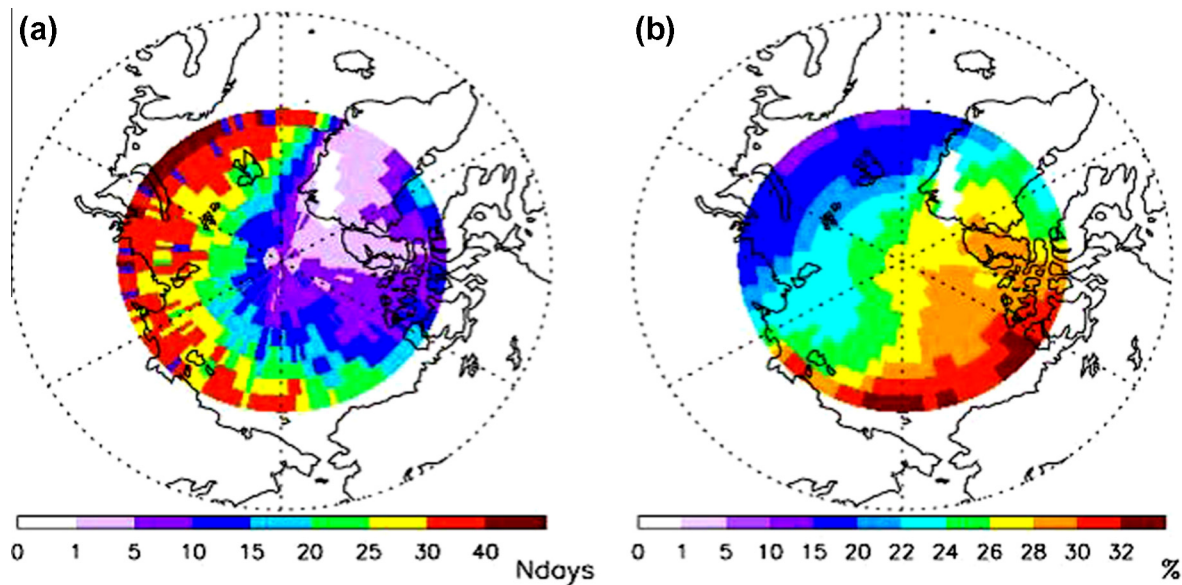


Fig. 5. Role of black carbon from wildfires in Arctic radiative forcing and haze. (a) Number of days during a 4-month period with optical depth greater than 0.094 (a characteristic value for Arctic haze events), and (b) contribution of the 2003 Russian fires to the optical depth of these days in percent (from Generoso et al., 2007).



fect the atmosphere through heat, water, and momentum exchanges.

## 4. Fire prediction

### 4.1. Burned area

Burned area by fires is one of the factors for fire emissions together with fuel loading, combustion efficiency, and emission factor. This information has been obtained using satellite detection and modeling approaches. Satellite remote sensing (RS) has emerged as a useful technique for fire and fuel mapping and monitoring. With the unique features of global coverage, high-resolution, and continuous operation, RS is able to obtain detailed information of fire occurrence, extent, structure, and temporal variation, together with the related fuel properties. Satellite instruments such as the Advanced Very High Resolution Radiometer (AVHRR) (Kaufman, 1990; Justice et al., 1996; Li et al., 1997; Burgan et al., 1998), the Geostationary Operational Environmental Satellite (GOES) (Prins and Menzel, 1990), and the Moderate Resolution Imaging Spectroradiometer (MODIS) (Kaufman and Justice, 1998; Justice et al., 2002) have been applied to active fire detection and burned area estimate. Giglio et al. (2010) developed a set of GFED3 burned area at a resolution of 0.5 degree, which was used for fire emission calculation in GFED3 (van der Werf et al., 2010). The MODIS burned area training data for a majority of grid cells, the MODIS 500-m daily burned area maps (Giglio et al., 2009), and a local regression approach were used. With the data set it was estimated that the global annual area burned for the years during 1997–2008 was 330–431 Mha, with the maximum occurring in 1998. The burned area varies to certain extent among various estimates.

There is large uncertainty in the global burned area, which can affect estimates of fire emissions and radiative forcing. According to a comparison with three other datasets for the period of 2001–2006 (Giglio et al., 2010), the GFED3 burned area was close to the 500-m MODIS burned area product (MCD45A1) (Roy et al., 2008), while the 1-km GLOBECARBON burned area product derived from SPOT VEGETATION, Along-Track Scanning Radiometer (ATSR-2), and Advanced ATSR (AATSR) imagery using a combination of mapping algorithms (Plummer et al., 2006) was close to the 1-km L3JRC product generated from Terra and Aqua MODIS imagery (Tansey et al., 2008). The L3JRC burned area was consistently much larger than all other data sets in about half of the regions examined, and consistently much lower than the GFED3 and MCD45A1 products in NH and SH Africa, and consistently overestimated the area burned in the continental United States and Canada each year by a factor of 3–10.

Fig. 6 shows the monitored burned area for the United States during 1960–2011 from the US National Interagency Fire Center (MIFC, [www.nifc.gov/fireInfo/fireInfo\\_statistics.html](http://www.nifc.gov/fireInfo/fireInfo_statistics.html)) and for Canada during 1970–2008 from the Canadian Wildland Fire Information System/Canada National Fire Database ([www.cwfis.cfs.nrcan.gc.ca/en\\_CA/nfdb/poly](http://www.cwfis.cfs.nrcan.gc.ca/en_CA/nfdb/poly)). The averaged burned area over the respective data period was 4.23 million acres (1.71 million ha) for the US and 2.1 million ha for Canada. The burned area over the period of 1997–2008 was about 2.43 Mha for the US and 1.89 Mha for Canada, with a sum of 4.32 Mha. The averaged burned areas over the period of 1997–2008 from the GFED3 (Giglio et al., 2010) were 1.5 for the Temperate North America and 2.2 Mha for Boreal and North America, which includes Alaska. The sum was 3.7 Mha. The GFED3 estimate is nearly 15% smaller than the monitored burned area for the temperate and boreal North America.

Fire models have been developed, which can be used not only to estimate historical fire activity but also project future fire trends

under changing climate. Fig. 7 shows the general structure of a fire model, which consists of schemes for fire occurrence and spread. Occurrence is dependent on how much fuel is available for burn, if there is an ignition mechanism (natural such as lightning or human such as arson), and how much risk of fuel and weather conditions for a fire to spread from the ignition location to the surrounding areas (weather and fuel conditions). Fire spread is characterized by its rate determined by fuel, weather and topography, and duration determined by fuel and weather condition and termination mechanisms such as fire wall and suppression.

Arora and Boer (2005) calculated the probability of fire occurrence based on the probability of fire ignition due to both lightning and human causes, the probability of fire spread that is dependent on wind direction and speed, and the probability of fire duration (extinguishing) determined by natural barriers and fire suppression. Pechony and Shindell (2009) estimated fire counts per month per unit square kilometer based on flammability (determined by vapor pressure, temperature, and precipitation, and vegetation), ignition, and suppression. Cloud-to-ground flash rates for lightning ignition were estimated using the polynomial regression with atmospheric thermal and microphysics parameters such as the convective available potential energy (CAPE) and the upward matter flux (UMF). The regression relations were valued using the LIS/OTD measurements (Christian et al., 2003). The human factor used population density as a major parameter. Using this model, Pechony and Shindell (2010) projected a shift to a global fire regime in the 21st century, suggesting that the future climate will possibly play a considerably stronger role in driving global fire trends, outweighing direct human influence on fire, a reversal from the situation during the last two centuries.

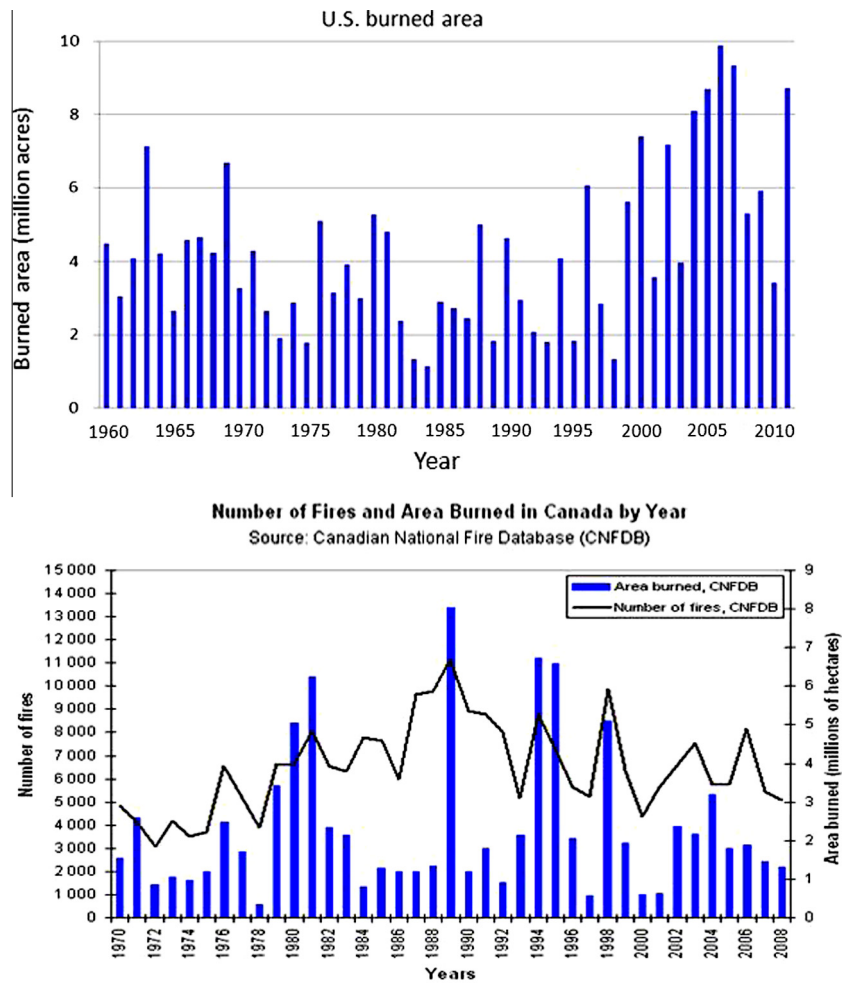
### 4.2. Fire risk

Fire risk indices are often analyzed and projected using fire weather indices and atmospheric and oceanic patterns that are associated with severe wildfire occurrence. Future fire potential trends can be determined by the differences in fire risk indices between now and the future. Most often used fire indices include the Keetch–Byram Drought Index (KBDI) (Keetch and Byram, 1968), Fire Weather Index (FWI) (Van Wagner, 1987), the Fosberg Fire Weather Index and its modified version (Fosberg, 1978; Goodrick, 2002), the Australian McArthur Forest Fire Danger Index Canadian (Luke and McArthur, 1978), energy release component (ERC), and burning index (BI). They are parts of fire danger rating systems such as the US National Fire Danger Rating System (NFDRS) (Deeming et al., 1977; Burgan, 1988) and the Canadian Forest Fire Danger Rating System (CFFDRS).

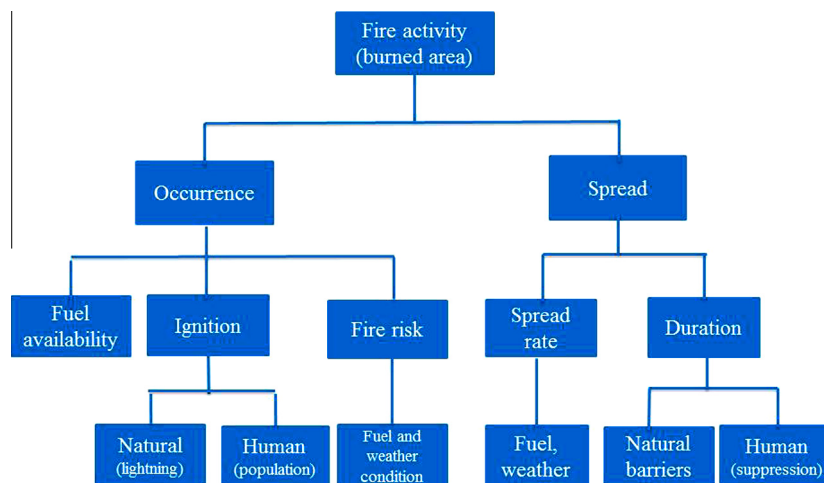
Among the three factors for fire occurrence, fire risk is most closely related to climate variability and climate change. It is also an important factor for fire spread. Thus, fire risk is a useful quantity to qualitatively measure future fire trends. Fire risk can also be directly related to burned area using statistical relations. Preisler and Westerling (2007) used a two-step method to estimate probability of certain large burned areas for a given 1-degree grid cell during a given month: probability of occurrence of at least one fire (ignition), and probability of fire spread or escape. Logistic regression with piecewise polynomials is used with factors including the Palmer Drought Severity Index.

### 4.3. Seasonal fire variability and oceanic conditions

Certain climatic patterns are strongly associated with prolonged periods of hot, dry, and windy conditions favorable for wildfire occurrence and spread. The El Niño Southern Oscillation (ENSO) is perhaps the best known climatic pattern impacting conditions across the United States. There are several other modes of climate



**Fig. 6.** Historical burned area in the United States and Canada. The panel for the US is plotted using the data obtained from the US National Interagency Fire Center (MIFC, [www.nifc.gov/fireInfo/fireInfo\\_statistics.html](http://www.nifc.gov/fireInfo/fireInfo_statistics.html)). The panel for Canada is from the Canadian Wildland Fire Information System/Canada National Fire Database ([www.cwffis.cfs.nrcan.gc.ca/en\\_CA/nfdb/poly](http://www.cwffis.cfs.nrcan.gc.ca/en_CA/nfdb/poly)).



**Fig. 7.** Diagram of a wildfire prediction model.

variability that can significantly influence regional climate. Teleconnection patterns are modes of atmospheric variability often revealed through empirical orthogonal function (EOF) analysis of geopotential heights and describe how different parts of the globe

are connected through the climate system. The relative importance of each teleconnection pattern typically varies with season, as some are dominant in the winter whereas others may dominate during the summer. The influences of several teleconnection

patterns manifest themselves by either enhancing or mitigating the impact of ENSO.

Studies on fire–climate relationships have revealed complex relationships that depend on the interaction of several teleconnection indices. For the western US, prime conditions for wildfires occur when abnormally wet years are immediately followed by drought and when ENSO transits from positive to negative phase (Swetnam and Betancourt, 1998; Kitzberger et al., 2001; Norman and Taylor, 2003). In the southeastern US, ENSO is also a dominant factor influencing wildfires. The dry conditions that usually occur during the negative phase of ENSO are the most critical as vegetation is abundant most years due to the generally moister conditions (Simard et al., 1985; Brenner, 1991). Heilman (1995) employed EOF analysis to identify atmospheric circulation patterns that were prevalent at the onset of severe wildfires across the United States. The two leading modes of this analysis represented the positive and negative phases of the Pacific–North American teleconnection pattern (PNA). Liu (2006) identified the closely coupled spatial patterns between Pacific SST anomalies and intense US fires using the singular value decomposition technique. Goodrick and Hanley (2009) provided evidence that the fire conditions in Florida were more complex than that described by Brenner (1991) as the variability in area burned is more explained by interactions between ENSO and the PNA than by climate pattern alone (Fig. 8). Dixon et al. (2008) reveal an even more complicated set of interactions in their investigation into Mississippi's wildfire activity that includes ENSO and the PNA as well as the Pacific Decadal and North Atlantic Oscillations (PDO and NAO respectively).

#### 4.4. Vegetation and fire

Fuel is a factor for fire occurrence and spread. Fuel conditions such as moisture, loading, and structure and geometry can significantly impact fire behavior and properties. Fuel conditions are expected to change under a changing climate (e.g., Zhang et al., 2010). Also, change and/or shift from a species type to another at a specific location, and in the composition of vegetation (e.g. mixture of tree species) or vegetation form, forest to woodland or woodland to forest can happen under a changing climate. Thus, climate change can impact fires indirectly through changing fuel conditions and vegetation types. Vegetation models including dynamic global vegetation models (DGVMs) have been used to predict future vegetation conditions and the related fire change. DGVMs are highly integrated process-based terrestrial ecosystem models that simulate daily or monthly carbon, water and nitrogen cycles driven by the changes in atmospheric chemistry including

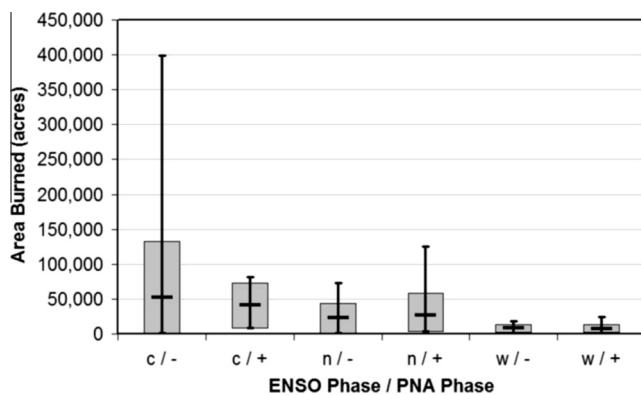


Fig. 8. Influence of ENSO and PNA teleconnection on Florida area burned. ENSO phase (c = negative, n = neutral, and w = positive) and PNA phase (– = negative and + = positive). Box represents  $\pm 1$  standard deviation and whiskers extend from minimum to maximum acres burned (from Goodrick and Hanley, 2009).

ozone, nitrogen deposition, CO<sub>2</sub> concentration, climate, land-use and land-cover types and disturbances. DGVMs usually include the core components of biophysics, plant physiology, soil biogeochemistry, and dynamic vegetation and land-use. The examples of DGVMs include HYBRIDS (Friend et al., 1997), MC1 (Bachelet et al., 2001), LPJ (Sitch et al., 2003), CLM (Levis et al., 2004), IBIS (Foley et al., 2005), and DLEM (Tian et al., 2010).

Fire is one of the disturbances included in these models. It is described either as an internal process or as an external forcing. As an internal process, it could change due to climate change expressed as varied boundary conditions of both atmospheric CO<sub>2</sub> and meteorological conditions (radiation, temperature, humidity, precipitation, etc.). The fire module in MC1 (Lenihan et al., 1998) simulates the occurrence, behavior, and effects of fire. The module consists of several mechanistic fire behavior and effect functions (Rothermel, 1972; Peterson and Ryan, 1986; Van Wagner, 1993; Keane et al., 1997) embedded in a structure that provides two-way interactions with the biogeography and biogeochemistry modules. The rate of fire spread and fire line intensity are the model estimates of fire behavior used to simulate fire occurrence and effects. The occurrence of a fire event is triggered by thresholds of fire spread, fine fuel flammability, and coarse woody fuel moisture.

## 5. Future fire trends under changing climate

### 5.1. Fire activities

Many studies have projected overall increases in burned area in boreal regions, but with varying magnitudes among the studies (e.g., Amiro et al., 2001; Flannigan et al., 2005; Balshi et al., 2009). An increase of 74–118% by the end of this century was obtained in a tripled CO<sub>2</sub> scenario (Flannigan et al., 2005). Liu et al. (2005) estimated an increase of 50% for the US average and over 100% for the western US by 2050. Spracklen et al. (2009) predicted that increases in temperature cause annual mean area burned in the western United States from 96 kha for the decade around 2000 to nearly 150 kha for the decade around 2050, an increase by 54%. Fig. 9 shows the increasing rates for individual regions in the western US. Krawchuk et al. (2009) used statistical generalized additive models (GAMs) to characterize current global fire patterns and projected the potential distribution of fire in the 21st century based on fire, climate, net primary productivity, and ignition data.

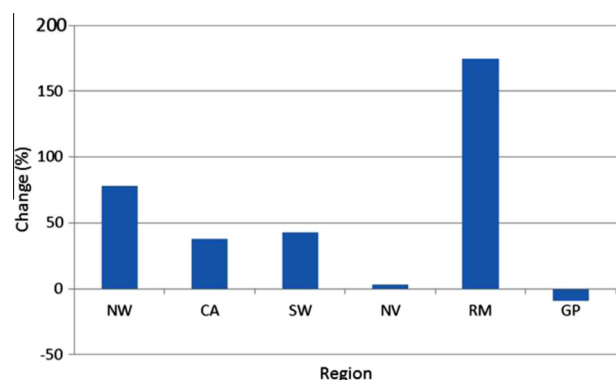


Fig. 9. Change rate (%) in burned areas of wildfires from present (1980–2004) to future (2046–2055) in the western United States. The regions are NW (Pacific Northwest), CA (California Coastal Shrub), SW (Desert Southwest), NV (Nevada Mountains/Semi-desert), RM (Rocky Mountain Forest), and GP (Eastern Rocky Mountain/Great Plains) (redrawing based on the results from Spracklen et al., 2009).

Lightning is projected to change under the changing climate, which is expected to affect fire ignition and therefore burned area. Wotton and Martell (2005) projected an increase of 80% in lightning initiated fire activity by the end of the 21st century. Prestemon et al. (2002) projected an increase of lightning initiated fires in the southern United States from about 43 kha in the year of 2002 to 124 kha by the mid-21 century. Empirical relations between fire activity and parameters including vegetation density, ambient meteorological conditions, availability of ignition sources, and fire suppression rates are used to project global fire trends based on the simulated climate variations and land-use changes.

### 5.2. Fire potential trends

A number of studies have provided more details about potential future North American and global fire trends using fire indices derived from regional or local climate change scenarios downscaled statistically or dynamically from GCM projections. Heilman et al. (1998) suggested the future occurrence of more surface pressure and atmospheric circulation patterns that are associated with severe wildfire occurrence in the eastern and southeastern US. Flannigan et al. (2001) used the FWI to show that future forest fire danger is expected to increase across most of Canada. Brown et al. (2004) projected the number of days of high fire danger measured by ERC is expected to increase in the western US by the end of this century mainly in the northern Rockies, Great Basin and the Southwest – regions that have already experienced significant fire activity this century. Liu et al. (2010) projected global fire potential using the KBDI for future climate projections from four GCMs under various emissions scenarios. This study projected increases in fire potential for western North America, southern Europe, central Asia, and central South America, central South Africa, and parts of Australia. For the US (Liu et al., 2012) (Fig. 10), future fire potential

is predicted to increase significantly in the Rocky Mountains for all seasons and in the Southeast and Pacific coast during summer and fall, with an exception in the inter-Mountain region where KBDI decreases in winter and spring. The increase in KBDI is more than 100 in many regions, which could be large enough to change fire potential level from low to moderate or from moderate to high.

All atmospheric conditions (temperature, humidity, wind, clouds/precipitation, etc.) are expected to change in response to the increased atmospheric CO<sub>2</sub> concentration. Studies (e.g., Gillett et al., 2004; Flannigan et al., 2005; Balshi et al., 2009; Liu et al., 2010, 2012; Parisien et al., 2011) have indicated that future warming is a more important contributor to projected increasing fire potential than the change in precipitation. For the regions and seasons with noticeable KBDI increases (50 or more), the magnitude of KBDI change due to the change in temperature is much larger than that due to the change in precipitation. With the consideration of future changes in relative humidity and wind conditions, future change in fire potential is more remarkable in the southern than northern portion of the western United States (Liu et al., 2012).

### 5.3. Vegetation change and fire impacts

Scholze et al. (2006) estimated changes in global ecosystem processes due to climate change during the 21st century. Simulations with LPJ using multiple climate change scenarios showed forest shifts and change in wildfire frequency, with high risk of forest loss for Eurasia, eastern China, Canada, Central America, and Amazonia and forest extensions into the Arctic and semiarid savannas. More frequent wildfire would appear in Amazonia, the far north, and many semiarid regions. Gonzalez et al. (2010) classified global areas into vulnerability classes by examining the changes of ecosystems in response to observed changes of 20th century

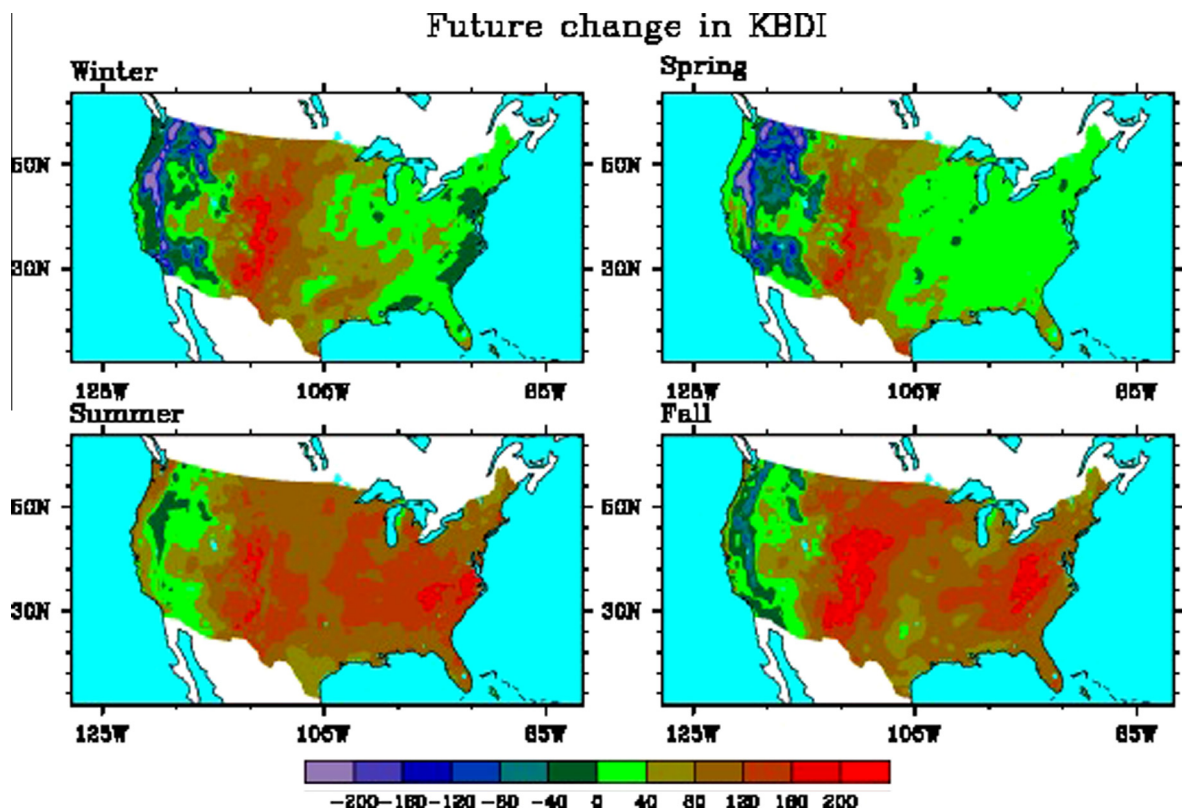


Fig. 10. Change in KBDI of North America for winter, spring, summer, and fall seasons between 2041–2070 and 1971–2000 calculated using the data obtained from the NARCCAP (from Liu et al., 2012).

climate and projected 21st-century vegetation changes using MC1. Temperate mixed forest, boreal conifer, and tundra and alpine biomes were found to have the highest vulnerability, often due to potential changes in wildfire. In addition, wildfire was projected to increase in the southeastern US, southeastern and northern China, and northern India.

Future changes in vegetation and fires were projected with MC1 for the US, especially western US and Alaska (Aber et al., 2001; Bachelet et al., 2001, 2005; Lenihan et al., 2003; Whitlock et al., 2003; Rogers et al., 2011). The simulated vegetation distribution (Bachelet et al., 2001) is dominated by broadleaf forest in Florida, Southeast mixed forest in the Gulf coast from east Texas to South Carolina, temperate deciduous forest in rest of Southeast, Northeast mixed forest in Northeast, and grassland/savanna/woodland in eastern Midwest. Future projection indicates an extension of Northeast mixed forest to eastern Midwest with the HadCM climate change scenario, and northward immigration of various forests for CGCM. Most significant changes in the western US are the disappearance of taiga/tundra in the Northwest and northern Rockies and the replacement of many arid lands by grassland in Southwest. The projected changes in regional vegetation patterns would significantly alter the occurrence and distribution of wildfires. The average annual acreage and biomass burned across the US is estimated to increase (Bachelet et al., 2003). By the end of the 21st century, 75–90% of the area simulated as tundra in Alaska is replaced by boreal and temperate forest (Fig. 11).

In addition to climate change, forest management could also modify future vegetation conditions and therefore affect wildfire. Amiro et al. (2001) pointed out that changes in fuel type need to be considered in a feedback of fire to global climate change (i.e., anthropogenic greenhouse gases stimulating fire activity through weather changes and fire releasing more carbon while the regenerating forest is a smaller carbon sink) because fire spreads more slowly through younger deciduous forests resulted from regeneration of burned forests. Fuel management options to reduce, convert, and/or isolate fuel can be a potential mechanism to reduce area burned. They can be used for small areas of the range of 100 kha, but are not realistic at the national scale of Canada. Furthermore, fire management agencies' ability to cope with the projected increases in future fire activity is limited, as these organizations operate with a narrow margin between success and failure; a disproportionate number of fires may escape initial attack under a warmer climate, resulting in an increase in area burned that will be much greater than the corresponding increase in fire weather severity (Flannigan et al., 2009b).

## 6. Conclusions and discussion

This paper has reviewed many of the studies on fire and climate interactions in the past few decades. The major findings from these studies include:

- Wildfire emissions can have remarkable impacts on radiative forcing. Smoke particles reduce overall solar radiation absorbed by the earth-atmosphere at local and/or regional scales during individual fire events or burning seasons. Fire emissions of CO<sub>2</sub>, on the other hand, are one of the important atmospheric CO<sub>2</sub> sources and contribute substantially to the global greenhouse effect.
- The radiative forcing of smoke particles can generate significant regional climate effects. It leads to a reduction in surface temperature. Smoke particles mostly suppress cloud and precipitation. Fire events could enhance climate anomalies such as droughts.
- Black carbon in smoke particles plays some different roles in affecting radiation and climate. BC could lead to warming in the middle and lower atmosphere, leading to a more stable atmosphere. BC also plays a key role in the smoke-snow feedback mechanism.
- Interannual variability in area burned is often related to ENSO and various teleconnection patterns. Unfortunately, climate models are limited in their ability to provide information on potential changes regarding ENSO variability and its interaction with various teleconnections in North America, which limits our ability to discuss future shifts in fire potential beyond just changes in the mean potential. However, the models are improving in this area and useful seasonal to multi-year projections of ENSO, AMO, etc., are probable in the next few years, which will improve prediction of interannual fire variability.
- Fires are expected to increase in many regions of the globe under a changing climate due to the greenhouse effect. Fire potential levels in the US are likely to increase in the Rockies all year long and in the Southeast during summer and fall seasons. Burned areas in the western US could increase by more than 50% by the middle of this century.

Many issues remain, which lead to uncertainties in our understanding of fire-climate interactions. Further studies are needed to begin to reduce these uncertainties. For fire particle emissions, a global picture of all kinds of radiative forcing is needed. It is a challenge considering the significant variability in both space and

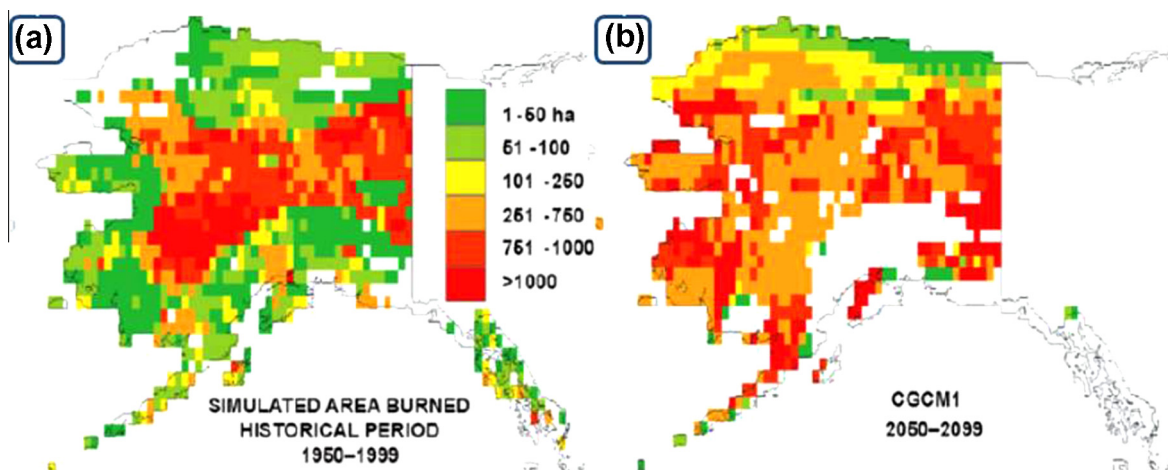


Fig. 11. Burned areas in Alaska (a) simulated for 1950–1999 and (b) predicted for 2050–2099 (from Bachelet et al., 2005).

time scales that characterize smoke emissions, along with the evolution of optical properties as smoke ages, and interactions with atmospheric dynamics and cloud microphysics. Smoke plume height and vertical profiles are important properties for impacts of smoke particles on the atmosphere, including locations of warming layers, stability structure, clouds, and smoke transport. Many simulation studies have been conducted based on assumed profiles. Some recent techniques such as the Multi-angle Imaging SpectroRadiometer (MISR) (e.g., Kahn et al., 2008) could be useful tools to determine these smoke plume properties. BC has received increased attention recently. BC emissions from fires, including emission factors from different fuels, need to be improved. In addition, BC and OC have different optical properties and climate effects. New techniques for measurement, analysis, and modeling are required to help investigate their separate and combined roles.

Work remains to be done on the assessment of the greenhouse effects and climate change deriving from fire CO<sub>2</sub> emissions. Unlike atmospheric total CO<sub>2</sub> concentrations, which have increased relatively steadily since the industrial revolution, fires have significant temporal variability. Fire regimes of a specific region may change dramatically as a result of changes in both climate and human activities. The variability can occur also over a short period. For example, the global carbon emissions in 1998 were 0.8 Pg C yr<sup>-1</sup> more than the average, but by 2001 they had dropped to 0.4 Pg C yr<sup>-1</sup> below the average (van der Werf et al., 2010). Thus, it is hard to estimate historical fire CO<sub>2</sub> emissions and their impacts. Furthermore, the contribution of wildfire emissions to global atmospheric CO<sub>2</sub> increase is more significant over a short period because regrowth of burned lands over a long period will remove some CO<sub>2</sub> from the atmosphere.

Many indices have been developed to measure fire risk, which is one of the factors for fire occurrence and spread, which is most closely related to climate change. More efforts are needed to build quantitative relationships with actual fire properties such as burned area. Although wildfires occur at local or regional scales, current climate models do not have the capacity to provide consistent and reliable simulation of climate variability at these scales, in particular for precipitation. The risk from mega-fires, which are small probability events and involve complex atmospheric, fuel, and human processes, would become larger under the projected warming climate. Many statistical climate–fire relations and vegetation models have very limited prediction skills for mega-fires. Fuel conditions such as type, loading, and moisture could change at a specific location in response to climate change. They will be also affected by human factors such as urbanization. Comprehensive approaches combining natural and social factors are needed for improving future fire projections.

While the strong relationships among atmospheric teleconnection/SST patterns and wildfire activity are useful for seasonal forecasting applications, their application to climate change scenarios is problematic. Joseph and Nigam (2006) revealed that the climate models used in the IPCC's Fourth Assessment report currently do a poor job simulating many features of ENSO variability and its interaction with various teleconnections in North America. ENSO–fire relations are valuable for seasonal fire predictions. USDA Forest Service and US National Oceanic and Atmospheric Administration joined research forces recently to develop plans and tools to improve fire weather and climate prediction skills, including exploring the SST–fire relations.

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