High latitude cooling associated with landscape changes from fires in North American boreal forests

B. M. Rogers¹, J. T. Randerson¹, and G. B. Bonan²

[1] {Department of Earth System Science, University of California, Irvine, California, USA; bmrogers@uci.edu, jranders@uci.edu}

[2] {Terrestrial Sciences Section, Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, Colorado, USA; bonan@ucar.edu}

Abstract

Fires in the boreal forests of North America are generally severe, killing the majority of trees and initiating succession that may last over a century. Functional variation during succession can affect local surface energy budgets and, potentially, regional climate with a change to long-term fire regimes. Burn area across Alaska and Canada has increased in the last few decades, and is projected to be substantially higher by the end of the 21st century due to a warmer climate with longer growing seasons. Here we simulated the impacts of augmented long-term fire regimes on boreal forest vegetation using an offline model of fire occurrence, historical fire data from national inventories, and succession trajectories derived from remote sensing. When coupled to an earth system model, younger vegetation from increased burning cooled the high-latitude atmosphere, primarily in the winter and spring, with noticeable feedbacks from the ocean and sea ice. Results from multiple scenarios suggest a relationship between surface air temperature and burning of \(-0.12 \pm 0.05^\circ\text{C}\), \(-0.24 \pm 0.09^\circ\text{C}\), and \(-0.43 \pm 0.13^\circ\text{C}\) per unit historical burn area increase for February - April, winter - spring, and annual time periods. This could provide a potentially important negative feedback to high-latitude terrestrial warming on the order of 2 - 4% and 11 - 18% annually, or 4 - 6% and 14 - 23% during winter, for a doubling and quadrupling of burn area, respectively. Further work is needed to integrate all the climate drivers from boreal forest fires, including aerosols and greenhouse gasses, and to incorporate Eurasia.
1 Introduction

Boreal forests represent one of the key biomes for understanding feedbacks to climate change because of their large area, carbon stocks (McGuire et al., 2002), biophysical imprints (Chapin et al., 2000; Euskirchen et al., 2010), and sensitivities to disturbance regimes (Weber and Flannigan, 1997; Lloyd et al., 2006). These systems experience particularly strong climate changes: northern high latitudes have been warming approximately five times faster than the global mean during the past few decades (Arctic Climate Impact Assessment, 2004), and are projected to warm by a further 3°C - 8°C by 2100 (Christensen et al., 2007). One of the primary feedbacks from boreal forests involves the effects of vegetation composition on surface energy budgets, which are amplified in these high-latitude environments by seasonal snow cover. In boreal North America, wildfire is currently the major agent of change for forest composition (Viereck, 1973; Payette, 1992; Chapin et al., 2006). These fires are typically severe enough to kill most living trees and initiate century-long vegetation succession that alters energy absorption and partitioning. Wildfire frequency and extent is controlled by summer high-pressure systems (Macias Fauria and Johnson, 2008) and patterns of temperature, precipitation, humidity, and wind (Flannigan and Harrington, 1988). Such conditions are projected to result in intensifications of boreal fire regimes during the 21st century. A comprehensive understanding of the potential impacts of altered fire regimes on vegetation distributions, energy budgets, and climate at the continental scale is thus critical for our understanding of high-latitude feedbacks to anthropogenic climate change.

Species and plant functional compositions of North American boreal forests are largely controlled by successional processes after fire. Mature forests (ie. several hundred years old) consist primarily of flammable evergreen conifers, including black spruce (Picea mariana), white spruce (Picea glauca), and to some extent in Eastern Canada, balsam fir (Abies balsamea) and white cedar (Thuja occidentalis). Immediately after a fire, winter and spring albedos are increased dramatically (up to 0.8) because of higher snow exposure, while on-site deposition of black carbon and charred surfaces reduce summer albedo (Chambers and Chapin, 2002; Amiro et al., 2006; Jin et al., 2012). Although this has a substantial impact on summer surface energy budgets, its persistence is limited to the first few years as black carbon is washed away and grasses, herbs, and
shrubs colonize and/or re-sprout (Amiro et al., 1999; Yoshikawa et al., 2003). Despite low albedos, net radiation and heat fluxes remain low because of decreased surface roughness and turbulent energy transfer (Chambers et al., 2005; Liu and Randerson, 2008). Re-growing landscapes during the first two decades are subsequently dominated by shrubs and tree saplings that exhibit higher summer albedos than pre-fire mature ecosystems. Winter and spring albedos continue to remain high, although they generally start to decline within a decade or two (Amiro et al., 2006; Beck et al., 2011).

Approximately 20 years after a fire, forest succession is dominated by trees. Two principal pathways are possible at the plant functional type (PFT) level: (1) self-replacement by evergreen conifers, or (2) relay floristics, in which deciduous broadleaf trees, primarily aspen (Populus tremuloides), birch (Betula spp.), or balsam poplar (Populus balsamifera), dominate and are gradually replaced by conifers after ~60 - 150 years. With the latter, summer albedo and latent heat remain higher, and sensible heat lower, than in mature forests because deciduous trees have brighter leaves and partition more of the available energy into transpiration during the growing season (Eugster et al., 2000; Liu and Randerson, 2008). Winter and spring albedos gradually decline as the canopy closes and/or deciduous trees, with near-zero leaf area outside the summer months, are replaced by conifers (Amiro et al., 2006; Lyons et al., 2008). Relay floristics therefore extend post-fire trends in albedo and transpiration for many decades, whereas conifer self-replacement expedites the return to pre-fire conditions. While studies have demonstrated that pre-fire forest age and composition, fire severity, climate zone, topography, aspect, and initial tree recruitment influence tree successional pathway (Mann and Plug, 1999; De Grandpré et al., 2000; Fastie et al., 2003; Johnstone and Kasischke, 2005; Johnstone and Chapin, 2006; Kurkowski et al., 2008; Johnstone et al., 2010; Beck et al., 2011), the prevalence of these across North America remains largely uncharacterized.

Boreal forest fires are projected to increase in frequency and severity by the end of the 21st century, due primarily to longer growing seasons and exacerbated mid-summer drought. Studies using fire weather indices or statistical relationships between burn area and climate variables from general circulation models generally predict increases in burn area on the order of 40% - 150% across Canada and Alaska by 2100.
(eg., Flannigan and Van Wagner, 1991; Stocks et al., 1998; Flannigan et al., 2005; Krawchuk et al., 2009; Amiro et al., 2009; Wotton et al., 2010). In an extreme case, Balshi et al. (2009) projected increases of 250% - 450%. Indeed, climate change-mediated intensifications of fire regimes are already being observed (Gillett et al., 2004): over the last three decades, burn area has been increasing in Canada (Podur et al., 2002; Stocks et al., 2002) and Alaska (Kasischke et al., 2010), with consequent increases in young deciduous stands (Bond-Lamberty et al., 2009). If fire regimes do follow such trajectories, we should expect a future landscape with younger stands and modified biophysics.

Recent research has identified the main drivers of fire regime-mediated biophysical impacts on high latitude climate in North American boreal forests. Field studies of fire chronosequences have documented typical post-fire trajectories of energy budget constituents, including albedo, net radiation, and fluxes of sensible, latent, and ground heat (Chambers and Chapin, 2002; Amiro et al., 2006; Liu and Randerson, 2008; McMillan and Goulden, 2008). Studies utilizing remote sensing have quantified trends of post-fire albedo (Lyons et al., 2008; Jin et al., 2012) and leaf habit (deciduous vs. evergreen, Beck et al., 2011) in various regions of Alaska and Canada. Randerson et al. (2006) demonstrated that a set of fires in Alaska resulted in globally-averaged radiative cooling. Euskirchen et al. (2009) used a dynamic vegetation and ecosystem model to project future fire regimes, PFT areas, and feedbacks to atmospheric heating in Alaska.

Here, we extend these analyses to the continental scale and simulate spatially- and temporally-explicit climate impacts within a coupled earth system model. Generalized succession curves are derived from remote sensing and the literature and used to drive vegetation distributions and climate under varying fire regimes. Results suggest a cooling trend with increased burning, strongest in late winter and early spring, that becomes amplified due to ocean and sea ice feedbacks after an approximate doubling of burn area.
2 Methods

2.1 Introduction

This research involved two primary undertakings: (1) the construction of driving datasets, including boreal forest extent, fire frequency, and vegetation succession, and (2) model simulations of the impacts of fire regimes on vegetation distributions, continental energy budgets, and high latitude climate. In the following sections we describe our approaches to these main tasks and to sensitivity analyses used to quantify uncertainty.

2.2 Driving datasets

2.2.1 Boreal forest domain

Our simulations were performed on a 1.9° latitude x 2.5° longitude (hereafter 2°) grid within the Community Earth System Model version 1.0.2 (CESM, Gent et al., 2011). CESM’s land model, the Community Land Model version 4 (CLM, Oleson et al., 2010), grid cells are composed of urban, glacier, lake, wetland, and natural vegetation fractions derived from the Moderate Resolution Imaging Spectroradiometer (MODIS), the Advanced Very High Resolution Radiometer (AVHRR), and long-term climate and crop datasets (Lawrence and Chase, 2007). Within natural vegetation, there are 17 plant functional types (PFTs), five of which occur in boreal/arctic North America. We defined our domain on this grid to be North American grid cells that contained at least 50% boreal/arctic vegetation cover within the vegetated fraction, at least 1% tree cover from MODIS, and some level of fire occurrence within the Alaskan Large Fire Database (Kasischke et al., 2002) or the Canadian National Fire Database (Amiro et al., 2001; Stocks et al., 2002; Parisien et al., 2006) (NFDBs) (Fig 1). Based on analyses of remote sensing data (below), within each grid cell a fraction of boreal/arctic vegetation was defined as boreal forest and the compliment as non-forest, or tundra. The tundra fraction was assigned a composition of boreal shrubs and arctic grass based on MODIS land cover over non-forest pixels, and the PFT composition of the boreal forest fraction changed as a function of our vegetation simulations (section 2.3.1).

A variety of land cover datasets can potentially distinguish boreal forests from tundra at moderate pixel resolution (~300m – 1km). These datasets are derived from classification schemes using high-resolution training datasets and vegetation indices from
satellite-based sensors, including AVHRR (e.g., Loveland et al., 2000; DeFries and Hansen, 2010; Ramankutty and Foley, 2010), MODIS (Friedl et al., 2002), the VEGETATION sensor (Bartholome and Belward, 2005), and the Medium Resolution Imaging Spectrometer (Bontemps et al., 2010). Initial analyses revealed discrepancies among and between datasets at field sites with known land cover, although MODIS performed best (data not shown). We therefore developed a pixel-level classification scheme based on MODIS land (MCD12Q1 year 2003) and tree cover (MOD44B) at 500-meter resolution to separate forest from tundra and other vegetation. Chronosequences over fire scar polygons from the NFDBs demonstrated that while MODIS land and tree cover displayed post-fire successional patterns, mean tree cover remained above 22% (Fig 2). 87% of unburned grass and shrub pixels contained tree cover under 18%, and 87% of burned pixels or unburned forest types (including savanna) contained tree cover over 18%. We therefore defined forest pixels as those with natural vegetation cover and tree cover ≥ 18% (Fig 1).

2.2.2 Historical fire regimes

Mean historical boreal forest fire return intervals (FRIs) were quantified at 2° using point versions of the NFDBs (Fig 1). FRI is defined here as the length of time required for the boreal forest fraction of a grid cell to completely burn, assuming no spatial overlap. Because burn area prior to 1960 is likely underestimated due to incomplete statistics (Amiro et al., 2001), we used data from 1961 – 2010 and accounted for missing years in particular Canadian provinces. Fires were distributed into CLM grid cells by assuming square-shaped burn areas on the landscape. Where concerned, temperate and boreal/arctic vegetation were assigned equal relative burn fractions. Forest pixels accounted for 85% of burn scars across the domain. Non-temperate burn area was therefore partitioned into boreal forest and tundra, scaled within each grid cell so that tundra pixels contributed to 15% of total burn area. Assuming a maximum of 1000 years, boreal forest FRI within a given grid cell was then calculated as

$$FRI = \text{MIN} \left( \frac{\text{boreal forest area (km}^2\text{)}}{\text{mean annual boreal forest burn area (km}^2\text{ yr}^{-1}\text{)}}, 1000 \right)$$

(1)
Empirical evidence suggests that younger boreal forests are less likely to burn due to lower levels of fine fuels and higher fractions of relatively non-flammable deciduous trees (Schimmel and Granström, 1997; Cumming, 2001; Lavoie, 2004). To account for this, we quantified age-dependent burn probabilities from fire scar polygons in the NFDBs. Ages of post-fire stands were tracked annually to calculate the cumulative probability of re-burn. A linear regression was fit to the annual probabilities for years 1 – 60 after a fire; thereafter, burn probability was assumed constant (Fig 2).

### 2.2.3 Succession trajectories

Post-fire succession, even in America’s boreal forests of low species diversity and high fire severity, is neither deterministic nor spatially constant (Bonan and Shugart, 1989). Re-vegetation involves complex interactions between fire severity, mortality, reproductive strategies, local seed sources, soil attributes, topography, and micro-climates (Viereck, 1973, 1983; Chapin et al., 2006; Johnstone et al., 2010b). Because of its long-term nature, there exists little quantitative information on succession patterns across large spatial domains. In this study, we attempted to characterize the mean patterns of succession at the PFT level across boreal North America. The ultimate goal was to represent surface energy budget anomalies via fire-mediated changes to vegetation distributions.

We used remote sensing analyses and expert opinion to construct post-fire vegetation succession trajectories (Fig 2). Because they displayed clear patterns and directly translate into CLM PFTs, we quantified evergreen needleleaf (EN) tree and shrub succession from chronosequences of MODIS land cover over fire scars from the NFDBs. A logistic curve was fit to the EN data, and a Weibull curve to the shrub. Deciduous broadleaf trees were assumed to dominate during mid-succession (~20 – 60 years), resulting in a pattern that qualitatively mimics 'relay floristics' succession. Freshly burnt land was prescribed a mixture of non-vegetated surfaces and herbaceous plants, represented by grasses. To account for the effect of charred surfaces on summer energy budgets, we assigned part of the non-vegetated fraction to a new PFT entitled ‘char’ (section 2.3.2). Although these trajectories are necessary over-simplifications, the validity of applying them to post-fire landscapes across boreal North America was tested in several ways (section 2.4).
2.3 Simulations

2.3.1 Vegetation

A simple model of fire occurrence was developed to quantify spatially-explicit boreal forest vegetation as a function of fire regime. Within each 2° CLM grid cell, boreal forest fractions were divided into sub-grid cells to represent the size of an average fire. Using fires 200 ha or larger from the point version NFDBs, this was approximated as 7,000 ha. Although these fires only represent 2-3% of the total number of fires in Canada, they account for over 97% of the burn area (Amiro et al., 2001; Stocks et al., 2002). Most grid cells contained a few hundred sub-grid cells.

The model runs on an annual time step. Within a given sub-grid cell, stand age is tracked and reset to zero when burning occurs. PFT composition then follows our succession curves until subsequent burning. The probability of burning is given by the product of a general 'fire seed', which is identical for every sub-grid cell, and stand age via the age-dependent fire probability function (section 2.2.2). At equilibrium, defined as $3\times10^4$ years, the mean FRI is calculated and compared to the desired FRI. The fire seed is adjusted until the two fall within 0.2% of each other.

For the simulation of present-day vegetation, the desired FRIs equaled those calculated for the historical period (BAx1, section 2.2.2). FRIs corresponding to multipliers of 0, 2, and 4 on historical burn area (BAx0, BAx2, and BAx4) were then used to quantify vegetation distributions as a function of altered fire regimes. We effectively applied a multiplier of zero by setting PFTs equal to maximum age compositions from our PFT trajectories.

2.3.2 Land model

We used the CLM version 4 with satellite phenology (CLM-SP) to quantify the impacts of vegetation on surface energy budgets. This version of the CLM incorporates remote sensing-based estimates of vegetation phenology to simulate fluxes of energy, momentum, and water. In order to better represent early post-fire water fluxes, we added a multiplicative scalar of 0.05 to bare soil evaporation in non-vegetated surfaces. This was applied only to the fraction of non-vegetated surface that resulted from burning. Recent parameterizations have addressed the CLM’s high evaporation bias over bare soil
(Oleson et al., 2008; Lawrence et al., 2010), although it remains an outstanding issue (P. Lawrence, pers. comm.). The bias was likely worse here due to the presence of organic matter in freshly burnt surfaces, which is assumed to be absent in the CLM’s non-vegetated PFT. Additionally, we added a new PFT, ‘char’, to capture the effect of charred surfaces and deposited ash on summer albedo. Char was represented by a plant with zero leaf area, small stature (0.2m), and low reflectance (visible and near infrared = 0.09 and 0.2, respectively, compared to 0.16 and 0.39 for tree and shrub stems, and 0.31 and 0.53 for grass stems).

Offline (uncoupled) CLM simulations were conducted for 40 years, driven by our modified vegetation distributions and repeating year-2000 climate and associated forcings. Although certain aspects of the model take years to reach equilibrium, especially soil water, the CLM displays no internal variability other than that inherited from external forcings. We therefore extracted year 40 for our analyses.

2.3.3 Climate model

The CESM includes interactive land, atmosphere, ocean, ice sheet, and sea ice models (Gent et al., 2011). Multiple configurations are possible in which certain components are inactive and replaced with observation-derived data. For our purposes, we ran CESM with active land (CLM-SP), atmosphere (CAM5), sea ice (CICE4), and a 'slab ocean' model (Bailey et al., 2011). The slab ocean simulates sea surface temperatures by using the mixed layer heat capacity and prescribing horizontal and vertical heat transport consistent with a fully coupled simulation. This allows for feedbacks between the atmosphere, ocean surface, and sea ice without adding the computational expense of the full three-dimensional model.

CESM was spun-up with repeating year 2000 forcings (biomass burning, volcanic, and anthropogenic aerosol emissions, solar radiation, and atmospheric CO2 concentration) for 35 years using land cover from our historical (BAx1) vegetation scenario. Four branch runs were then initiated with land surfaces corresponding to BAx0, BAx1, BAx2, and BAx4. Simulations ran for an additional 120 years with the exception of BAx4, which ran for only 80 years because of its stronger forcing and climate responses.
2.4 Validation

Our generalized PFT curves, fire regimes, and land model forcings were validated in several ways. Firstly, our succession trajectories were input to the CLM to simulate post-fire energy budgets. A number of variables were compared to field- and satellite-based observational datasets, including Chambers and Chapin (2002) (summer albedo, net radiation, latent heat, and sensible heat), Liu et al. (2005) (ground temperature), Amiro et al. (2006) (winter albedo and summer net radiation, sensible heat, latent heat, Bowen ratio, albedo, and ground heat flux), Randerson et al. (2006) (monthly albedo), Liu and Randerson (2008) (net radiation, sensible heat, and latent heat), Lyons et al. (2008) (spring and summer albedo), and McMillan and Goulden (2008) (summer albedo). In all cases, the CLM was run offline under repeating year 2000 climate for 20 years, after which the PFT-succession curves were implemented. The grid cell(s) and month(s) containing the observations was(were) extracted for comparisons.

Boreal vegetation distributions are functions of fire frequency and PFT succession from our vegetation model (section 2.3.1), and MODIS land cover over tundra pixels. This assumes post-fire succession to be the only process controlling vegetation composition in boreal forests. Although a variety of factors affect vegetation in boreal North America, including climate, solar radiation, soil moisture and temperature, nutrient availability, permafrost, depth of forest floor organic layer, proximity to seed sources, reproductive legacies, and other disturbances (Heinselman, 1978; West et al., 1981; Van Cleve et al., 1986; Bonan, 1989; Bonan and Shugart, 1989; Soja et al., 2007), fire is arguably the most influential (Weber and Flannigan, 1997). To validate this approach, simulated present-day vegetation was compared to continental-to-global land cover datasets, including MCD12Q1, GlobCover 2009 (Bontemps et al., 2010), GLC2000 North America (Bartholome and Belward, 2005), University of Maryland Land Cover (DeFries and Hansen, 2010), ISLSCP II Potential Natural Vegetation Cover (Ramankutty and Foley, 2010), the Global Land Cover Characteristics Data Base Version 2.0 (Loveland et al., 2000), and the CLM input PFT distributions (Lawrence and Chase, 2007).
2.5 Sensitivity analyses

Four sensitivity studies were conducted in order to better characterize uncertainty. Due to the demanding computer resources of the coupled CESM, our analyses focused on energy budgets in the offline CLM. In each case, vegetation distributions from BAX1 and BAX2 were used, and the CLM was run for 40 years with repeating year 2000 forcings. In our main simulations, burn area increased spatially in direct proportion to historical distributions. Given that young stands are less likely to burn and spatial patterns of burning may change in the future, we implemented a scenario (‘Add BA’) in which burn area increased uniformly across boreal forests in the domain.

Analysis from the NFDB polygons revealed that nearly all young forest re-burns (<10 years) occurred near the perimeters of burn scars (data not shown), where mapping errors are suspected to be greatest. This suggests that the actual probability of re-burn in young stands may be smaller than our regression indicates (section 2.3.2). To address this, our second sensitivity analysis (‘PF zero’) used an age-dependent burn probability function that passed through the point (0,0), but was otherwise identical to the main regression.

Despite being validated in several ways, there remains a high level of uncertainty regarding our succession curves, particularly the contribution of deciduous broadleaf (DB) trees. In our main trajectories, the EN tree fraction was taken directly from a regression of MODIS EN tree pixels over fire scars. We assigned the non-EN tree fraction in mid-successional years exclusively to DB trees. However, in MODIS, this includes DB trees as well as mixed forests and savannas, which undoubtedly contain contributions from conifers. We therefore conducted simulations (DBx0.5) in which the DB fraction was halved, and the removed fraction assigned to EN trees.

In our main simulations, multipliers were applied to boreal forest burn area. There is, however, a negative feedback to changes in burn area due to altered vegetation and stand ages (Fig B) This biologic feedback has been neglected in most previous studies' projections of burn area (eg., Flannigan et al., 1998; Stocks et al., 1998; Amiro et al., 2009; Balshi et al., 2009; Krawchuk et al., 2009), which used various metrics of fire weather and fuel loads. We quantified this biologic feedback by conducting simulations (‘Bio Feedback’) in which multipliers were applied to the historical 'fire seed', a proxy for
the combination of historic fire weather and fuel dynamics, instead of burn area. This scenario was not implemented in the CLM, because only burn area changed and surface forcing variables were found to be linear from BAx0 - BAx2 in the main simulations (section 3.3).

3 Results

3.1 Validation

After minor parameterization changes and implementation of the ‘char’ PFT, CLM-simulated post-fire energy budgets compared favorably with observations (Fig 3). A higher level of uncertainty exists for both observations and model-observation comparisons during winter and spring. This is likely due to a greater dependence of winter and spring albedo on burn severity, snow cover, and time that dead boles remain standing. While comparisons with Amiro et al. (2006) imply a high bias in modeled post-fire winter/spring albedo, data from Lyons et al. (2008) suggest model estimates are high for the first decade and low thereafter. Comparisons of monthly albedo with Randerson et al. (2006) are favorable over the entire year for a mature and early burn site, as are latent and sensible heat (data not shown). Considering the variability and uncertainty inherent in post-fire succession and measurement techniques, we believe our approach captures the important dynamics of post-fire energy budgets.

Domain-wide vegetation fractions from remote-sensing based datasets display a high amount of variability (Table 1). This is to be expected for categorical data derived from different satellite sensors, band wavelengths, training layers, analysis years, vegetation classes, classification techniques, and experts’ interpretation. Nonetheless, our simulations of historical vegetation compared reasonably well. It must be noted that direct comparisons are not possible due to differences in vegetation classification: observational datasets include mixed forests, woodland/savanna, and/or tundra, all of which are not represented by distinct CLM PFTs. Although our simulated deciduous broadleaf tree coverage is higher than most datasets, deciduous stands are comparatively patchy on the landscape and likely comprise a substantial portion of the mixed forests and woodland/savanna classes.
3.2 Vegetation simulations

In our vegetation model, North American boreal forests varied from nearly pure evergreen needleleaf (EN) stands at long fire return intervals (FRIs) to mixtures of EN trees, deciduous broadleaf (DB) trees, shrubs, grasses, and non-vegetated cover at short FRIs (Fig 4). DB trees were more dominant than EN trees in stands with FRIs less than ~60 years, and peak at ~40% coverage in grid cells with FRIs of ~45 years. This agrees qualitatively with observations: across Canada, forest inventory data suggest that the deciduous fraction of forests varies from 1% to 40% depending on ecozone (Amiro et al., 2001).

Augmented fire regimes, represented by multipliers of 0, 2, and 4 on historical burn area (BAx0, BAx2, and BAx4), had a pronounced effect on simulated boreal forest vegetation across the domain (Fig 4). EN tree fraction was approximately halved from 62% in BAx0 to 32% in BAx4, with DB trees and shrubs accounting for most of converted vegetation. Including tundra pixels, shrub coverage nearly equaled EN tree coverage under BAx4 (28% shrubs and 32% EN trees).

3.2 Offline Land Model

The largest impacts on domain-wide CLM surface energy budgets occurred during the late winter and spring months (Fig 5). Increased burning resulted in higher albedos due to more snow exposure. The combination of higher albedos and increasing solar insolation and in the spring substantially reduced net radiation and sensible heat (~ -7.5% for BAx2, and ~ -17.5% for BAx4, for February – April across the domain).

Anomalies during summer months were comparatively smaller than those during late winter and spring (Fig 5). Summer energy budgets were affected by differences in partitioning and turbulent exchange. With increased burning, domain-wide albedos were elevated slightly. Although less solar radiation was absorbed, outgoing longwave was higher due to shorter roughness lengths. Both of these factors acted to decrease net radiation. However, because deciduous trees partition more available energy into transpiration during the growing season, latent heat was still elevated (+1.6% and +3.6% for BAx2 and BAx4, respectively, from June - August). Sensible heat was thus further reduced.
Fall and early-mid winter displayed minimal changes, as solar insolation and snow accumulation were relatively low. On an annual basis, more burning resulted in higher albedos, less net radiation and sensible heat, and slightly more latent heat (Table 2). We would expect this to result in cooling, strongest in winter and spring, and atmospheric moistening during the summer. However, local and remote responses and feedbacks, involving the atmosphere, ocean, and sea ice, determine the patterns and magnitudes of climate impacts from such forcings.

3.3 Coupled Climate Model

Late winter and spring temperatures were significantly affected by albedo changes (Table 3). The largest domain-mean anomalies occurred in March for BAx0 (+0.73°C), and February for BAx2 (-0.49°C) and BAx4 (-1.41°C) (Fig 6). Changes were generally greatest across the high-burning areas of Alaska, central Canada, and Quebec (Fig 7), reaching maximum grid cell values of +1.9°C (BAx0), -1.2°C (BAx2), and -3.0°C (BAx4) for mean February - April anomalies.

While surface forcings displayed a dampened response to increasing burn area, temperature anomalies were linear, if not amplified (Fig 8). A number of feedbacks were responsible for this. Prevailing westerlies relayed continental temperature anomalies to the North Atlantic and Greenland (Fig 7), cooling the sea surface and increasing annual sea ice coverage (Fig 9): mainly polar ice in the summer and fall, and sub-polar ice in the winter and spring. This produced further cooling in eastern Canada. Snow coverage within the domain also increased (+2.6% and +8.1% in April - May for BAx2 and BAx4, respectively), amplifying the positive land surface albedo anomalies (Fig 5). Total column atmospheric water dropped slightly with increased burning (from +0.6% in BAx0 to -1.1% in BAx4 annual means), reducing its greenhouse effect. Regressions on the relatively linear responses from BAx0 – Bax2 yield slopes of -0.12, -0.24, and -0.43 °C/(unit historical burn area), or -0.06, -0.11, and -0.20 °C/(Mha yr⁻¹), for annual, winter – spring, and February – April time periods, respectively (Table 5).

As a result of climate feedbacks and, potentially, model biases, domain-wide surface energy flux anomalies were somewhat dissimilar between online and offline simulations (Table 2). While land surface albedo changes were comparable, the ground surface cooled considerably more with increased burning in the coupled model, which
tended to reduce longwave emission and dampen net radiation anomalies. Additionally, when coupled to cooler atmospheres with less evaporative demand, the land surface in BAXx2 and BAXx4 evaporated less water throughout the winter, spring, and early summer (Fig 5). Although surface specific humidity decreased during these months, so did temperature, so that spring vapor pressure deficits were reduced by ~3% in BAXx2 and ~7% in BAXx4 (Table 3). This was likely exacerbated in CESM due to a high temperature bias over the domain during these months (Fig 6), which would tend to amplify anomalies in vapor pressure deficits via the non-linear Clausius-Clapeyron relation. Latent heat changes were therefore opposite in sign in BAXx2 and BAXx4 between the two sets of simulations (Table 2, Fig 5), and sensible heat anomalies were in turn further dampened. As a result of this and other potential dynamical changes, precipitation displayed a weak negative trend with increased burning (Table 3).

The boreal atmosphere responded in other notable ways to surface forcings. Because of decreases in sensible heat associated with more burning, planetary boundary layers were lowered throughout the year, with greatest reductions in spring (Table 3). Cooling also created high surface pressure and anti-cyclonic wind anomalies in BAXx2 and BAXx4; the opposite was true for BAXx0. Wind anomalies were greatest in winter, when anti-cyclonic flow dominated, with domain-wide values of -1.3%, +1.5%, and +3.4% for BAXx0, BAXx2, and BAXx4, respectively. A number of atmospheric responses reached beyond the boreal domain into temperate North America and the North Atlantic (eg. Fig 7).

### 3.4 Uncertainties

Sources of uncertainty, bias, and fallible assumptions in a study such as this are numerous. In an effort to evaluate their impacts, we quantified two major sources: driving datasets and climate model responses. For the driving datasets, we conducted three sets of BAXx1 and BAXx2 offline land simulations with altered boundary conditions (Table 4): (1) burn area was added uniformly across the domain ('Add BA'), (2) the burn probability vs. stand age function was forced through the point (0,0) ('PF zero'), and (3) the DB tree composition throughout succession was halved ('0.5 DB'). As a fourth estimate ('Bio Feedback'), we assessed the biologic feedback to fire occurrence by multiplying our spatially-explicit historical fire probability seeds, instead of burn area, by factors of 2 and
Burn area was reduced by 21.7% and 30.5%, respectively. Because net radiation and surface temperatures responded relatively linearly from BAx0 - BAx2 (Figure 8, Table 2), we used the variance in net radiation changes, and burn area reduction from Bio Feedback, between BAx1 and BAx2 as an estimate of uncertainty in our driving datasets (Table 5). This was done separately for each seasonal period.

The CESM coupled climate model displays a considerable amount of high-latitude inter-annual and inter-decadal variability because of long-term oscillations in atmosphere-ocean dynamics (Jahn et al., 2011; Landrum et al., 2011). This is true even with the slab ocean model (Subin et al., 2012), and especially prominent in winter: we found the standard error of linearly de-trended surface temperature for land north of 45°N in CESM for BAx1 to be 2.37 °C vs. 1.97°C in Qian et al. (2006) reanalysis data. Due to computing resources, we were able to run our equilibrium climate simulations for a maximum of 120 years each, with the exception of BAx4, which ran for 80 years. To quantify uncertainty in our relationship of domain-wide temperature vs. burn area due to limited run time, we performed $1 \times 10^6$ Monte Carlo simulations of the estimated slope from BAx0 - BAx2 for each seasonal period. In each case, simple linear regressions were fit to three points chosen from the normally distributed populations of $\Delta$(temperature) (Fig 8).

These two major sources of uncertainty were pooled, and resulted in uncertainties of 41%, 37%, and 29% for slopes of temperature vs. burn area during annual, winter - spring, and February - April time periods (Table 5).

4 Discussion

Earth’s climate-vegetation system is especially sensitive to change at high latitudes. Thresholds of plant establishment and survival are crossed over large areas with changing temperatures. Once the potential for forest is realized, climate-dependent disturbance regimes place strong controls on species composition. Snow exposure amplifies albedo effects from structural vegetation changes. The relatively shallow boreal atmosphere responds sharply to these radiative perturbations, and additional feedbacks from glaciers, sea ice, and sea surface temperatures amplify the responses.
Researchers have explored the sensitivity of climate to boreal vegetation since the inception of credible land components within general circulation models. Otterman et al. (1984) calculated the high-latitude atmosphere to be 4.6°C cooler without the masking of snow from vegetation. Thomas and Rowntree (1992) demonstrated that boreal deforestation could increase land albedo and snow depth, and decrease surface temperature, net radiation, heat fluxes, and precipitation during spring. Bonan et al. (1992) estimated this cooling to be upwards of 12°C at ~60°N in April when amplified by ocean feedbacks. More recently, Betts (2000) calculated a +10 to 20 W m\(^{-2}\) surface shortwave forcing from afforestation in boreal regions. In a complementary deforestation experiment, Bala et al. (2007) estimated a 10.7% increase in albedo, -10.6 W m\(^{-2}\) surface shortwave forcing, and 2.1°C annual cooling in the 50°N - 90°N high latitude belt. This study attempts do draw upon this earlier work but use a data-driven approach to hone in on responses from a particularly influential and climate-sensitive land cover agent: boreal forest fires.

Our surface forcings are in general agreement with recent work in Alaska. Euskirchen et al. (2009) estimated an increase in summer albedo between 0.006 and 0.015, and winter albedo of 0.05, across the western arctic associated with burn area increases on the order of 100 – 200% under future climates. For the same domain, we estimate albedo increases of 0.003 and 0.005 for summer, and 0.021 and 0.057 for winter, for B Ax2 and B Ax4, respectively. Randerson et al. (2006) estimated a cumulative atmospheric radiative forcing of -4.2 W m\(^{-2}\) due to surface albedo over the first 80 years for one fire in interior Alaska. A reduction in fire return interval (FRI) from 80 to 40 years was estimated to cause a radiative forcing of ~ -1.9 W m\(^{-2}\). In the same location, we estimate surface forcings of approximately -6 W m\(^{-2}\) over the course of the first 80 years post-fire (-6.0 W m\(^{-2}\) for surface shortwave and -5.4 W m\(^{-2}\) for net radiation). A drop in FRI from 80 to 40 years produced a surface forcing of ~ -2 W m\(^{-2}\) (-2.4 W m\(^{-2}\) in surface shortwave and -1.9 W m\(^{-2}\) in net radiation).

Coupled model results suggest a near-linear response of land surface temperature to vegetation changes associated with a decrease or increase in burning up to ~100%, and somewhat dampened forcings, yet amplified responses, thereafter. Although offline simulations suggest greater summer evaporation with younger deciduous forests, cooler
temperatures in the coupled model reduce evaporative demand and lead to a slightly drier atmosphere with a shallower boundary layer. Implied is a future with higher temperatures, longer growing seasons, and increased burning. Assuming a 3 - 5°C increase in annual surface air temperature, and 5 - 8°C increase in winter, over boreal North America by 2100 (Arctic Climate Impact Assessment, 2004; Christensen et al., 2007), our results imply fire-mediated land surfaces could contribute a negative feedback of 2 - 4% and 11 - 18% annually, and 4 - 6% and 14 - 23% during winter, under a doubling and quadrupling of burn area, respectively.

It must be noted, however, that these responses and feedbacks will not occur independently. Our assumed increases in burning are themselves results of anthropogenic greenhouse warming. Ocean and sea ice feedbacks will respond non-linearly to an overall warming trend, and may therefore not interact with this smaller cooling footprint to the same degree. Additionally, the boreal and arctic land surface will likely respond to 21st century climate via mechanisms other than fire. With permafrost melting and surface air warming, we have witnessed increases in the northern boundaries and productivities of boreal shrubs during the last several decades (Sturm et al., 2001; Beck and Goetz, 2011), leading to an estimated +0.2 W m⁻² of atmospheric heating per decade since 1950 (Chapin et al., 2005). It has been estimated that shrubland coverage under future climates has the potential to add over 6 W m⁻² of atmospheric heating (Chapin et al., 2005), with additional warming coming from atmospheric water vapor feedbacks via shrub transpiration (Bonfils et al., 2012). Boreal forests, in contrast, have become less productive due to drought stress from higher temperatures and longer growing seasons (Beck and Goetz, 2011), and may take much longer than shrubs to migrate northward (Chapin and Starfield, 1997; Rupp et al., 2001).

Boreal fires also affect climate in ways other than land surface characteristics. Pyrogenic aerosols released during combustion interact with the atmosphere and planetary radiation balance in complex ways. Typically, direct effects from fire aerosols produce warming in the local atmosphere but cooling at the surface (Stone et al., 2008; Tosca et al., 2012). In the arctic, however, top of atmosphere energy balance depends on the albedo of the underlying surface (Ramanathan et al., 2001; Stone et al., 2008), and deposited black carbon can produce strong local warming and melting of snow and sea
ice (Flanner et al., 2007; Ramanathan and Carmichael, 2008). When indirect effects on cloud formation are considered, the atmosphere is further cooled and radiative forcing becomes strongly negative (Ward et al., 2012).

Fires also release greenhouse gases: mostly CO$_2$, but also CH$_4$ and N$_2$O, which are well-mixed and produce positive global radiative forcings. A 50% increase in circumpolar boreal forest fires has the potential to release 27 – 51 Pg C, or 0.33 – 0.8 Pg C yr$^{-1}$ during the next 50 – 100 years (Kasischke et al., 1995). Model estimates suggest that Canada’s fire emissions may increase considerably more than this, driven mostly by increasing burn area (Amiro et al., 2009). The overall biogeochemical forcing, however, must include ecosystem recovery rates, which are different for different carbon pools and depend on climate, fire severity, and species composition. Surface biophysical, aerosol, and biogeochemical effects from fires must all be considered when assessing the overall impacts of fires on climate. For one fire in Alaska, Randerson et al. (2006) found that surface albedo dominates the radiative budget, approximately doubling the other positive forcings in magnitude and leading to an overall negative radiative forcing. Aerosol indirect effects were not considered, however.

While we attempted to quantify several influential yet accessible uncertainties, a variety of potential biases and errors remain untreated. One outstanding issue involves post-fire succession. Simulated vegetation distributions were direct functions of succession, which is highly uncertain. Our approach supported the hypothesis that forests in boreal North America primarily re-grow through relay floristics, wherein deciduous trees prevail during mid-succession. Although this has a large potential for error, comparisons of energy budgets to observations were favorable, and halving the prescribed deciduous fraction had a relatively minor impact on energy budgets (Table 4). More pertinent for uncertainty, however, is the assumption that these trajectories are constant across space, time, and levels of fire severity. Observations suggest post-fire succession is influenced by geography and topography, and depends on pre-fire conditions. Specifically, higher proportions of deciduous trees establish and survive during succession on south facing slopes with improved drainage, and with greater pre-fire deciduous fractions, higher fire severity, and increased fire frequency. Deciduous forests are less flammable. Krawchuk and Cumming (2011) estimated a biotic feedback
on burn area of ~23% under a scenario that otherwise predicted an increase of 61%. This study estimates a 22% reduction on a x2 burn area prescription. However, it has been suggested that an intensifying fire regime may not just increase deciduous fractions, but lead to an alternate stability point of deciduous forest dominance (Johnstone et al., 2010). A landscape dominated by deciduous trees, instead of conifers and patches of mixed stands, may display entirely different fire dynamics.

Fire severity (carbon combusted per unit burn area) may also change in the future, altering mortality and succession. While some studies suggest an increase in severity due to climate (e.g., Harden et al., 2000; Rogers et al., 2011) and drier forest floor organic matter (OM) (Bonan et al., 1990), less flammable deciduous forests with increasingly smaller pools of OM may decrease severity. Additionally, we did not attempt to address overarching model biases in either the CLM land model or the CESM climate model. Other earth system models with different representations of energy and water fluxes, atmospheric circulation, sea ice dynamics, ocean fluxes, and climate sensitivity may display different quantitative and qualitative responses. Finally, we include no representation of time scale in our equilibrium simulations, although landscape changes are implicitly assumed to take place over one to two centuries.

Further work that expands the spatial domain, forcing agents, and modeling techniques considered here is needed to better clarify the impacts of boreal forest fires on climate and ecosystems over the next century. Eurasia must be considered, as it contains approximately twice the coverage of boreal forest as does North America, with distinct and heterogeneous ecology and fire regimes. Although surface forcings may dominate in high latitudes, ecosystem carbon budgets and fire emissions of aerosols and greenhouse gasses must be included in order to adequately represent the full climate impacts. With the continuing development of global land models, dynamic fire and vegetation modules should be used to capture the non-linear and interacting responses of succession, burn area, and fire severity to climate change. Additionally, analyses that better characterize spatial and temporal relationships between succession, fire intensity, burn severity, and surface forcings can provide better constraints.
4 Conclusions

Modeling results suggest that increases in fire activity across the North American boreal forest, similar to that projected by various research groups for the 21st century, would result in a younger landscape that cools the high latitude atmosphere. Amplified by ocean and sea ice feedbacks, impacts are dominant in February – April when solar insolation is moderate but albedo anomalies are large. Temperature anomalies decrease by a factor of ~2 for the 6-month winter/spring, and by a factor of ~4 for annual means. Annually, this would provide only a minor reduction to high-latitude climate warming with a doubling of historical burn area, but could decrease warming by nearly 25% in the winter with a quadrupling of burn area. Although this result implies a negative feedback to climate change, future work is necessary to adequately quantify the spatial and seasonally-varying effects from all fire forcing agents: landscape biophysical, aerosol emissions, and greenhouse gases. Particularly, while landscape and aerosol emissions likely cool the high-latitudes, pyrogenic greenhouse gases warm the global atmosphere.

This work also implies that representing fire and vegetation succession in earth system models is crucial in order to project high-latitude climate and ecosystem responses to anthropogenic forcings. Dynamic vegetation models that simulate distinct age, functional, and structural classes can help characterize contemporary succession in locations with limited data, and project how succession and migration will ultimately shape future distributions. These should be coupled to mechanistic fire models that respond to climate and represent spatially-varying fuel dynamics to simulate fires of different types and intensities. Until such models are validated against historical and paleo archives and fully integrated into climate models, we can only provide boundary conditions for possible future impacts.

Finally, while this study suggests that boreal forest fires can provide regional negative feedbacks to warming, it must be noted that fires are associated with a number of deleterious non-climate impacts. Forest fires decrease air quality and damage homes and tourism industries throughout the world. While regular low-intensity fires may increase the health of temperate forests by reducing ground cover and ladder fuels and increasing biodiversity, this is not necessarily the case for boreal forests. Monospecific
boreal stands display minimal structural layers, and regional biodiversity is low and maintained through climate and edaphic gradients.
References


Bergeron, Y.: Species and stand dynamics in the mixed woods of Quebec’s southern boreal forest, Ecology, 81(6), 1500–1516, 2000.


Subin, Z. M., Murphy, L. N., Li, F., Bonfils, C. and Riley, W. J.: Boreal lakes moderate seasonal and diurnal temperature variation and perturb atmospheric circulation: analyses in the Community Earth System Model 1 (CESM1), Tellus A, 64(0), 2012.


Table 1. Domain-wide vegetation types for historical offline vegetation simulations and global land cover datasets.

<table>
<thead>
<tr>
<th>Vegetation Type</th>
<th>BAX1</th>
<th>CLM</th>
<th>MODIS</th>
<th>GlobCover 2009</th>
<th>GLC2000</th>
<th>UMD</th>
<th>Ramankutty &amp; Foley</th>
<th>IGBP</th>
<th>USGS</th>
</tr>
</thead>
<tbody>
<tr>
<td>EN tree</td>
<td>49.4</td>
<td>45.0</td>
<td>23.4</td>
<td>36.5</td>
<td>40.3</td>
<td>23.9</td>
<td>45.1</td>
<td>20.5</td>
<td>20.5</td>
</tr>
<tr>
<td>DB tree</td>
<td>7.9</td>
<td>3.2</td>
<td>0.04</td>
<td>1.6</td>
<td>1.3</td>
<td>0.3</td>
<td>0.2</td>
<td>0.3</td>
<td>8.5</td>
</tr>
<tr>
<td>Shrub</td>
<td>22.4</td>
<td>23.3</td>
<td>25.9</td>
<td>1.0</td>
<td>11.4</td>
<td>19.9</td>
<td>0.04</td>
<td>39.3</td>
<td>8.5</td>
</tr>
<tr>
<td>Grass</td>
<td>2.4</td>
<td>10.9</td>
<td>2.0</td>
<td>6.7</td>
<td>0.3</td>
<td>3.9</td>
<td>0.9</td>
<td>0.9</td>
<td>0.9</td>
</tr>
<tr>
<td>Not-vegetated</td>
<td>7.5</td>
<td>7.3</td>
<td>0.04</td>
<td>0.3</td>
<td>1.8</td>
<td>2.5</td>
<td>1.6</td>
<td>1.8</td>
<td>0.8</td>
</tr>
<tr>
<td>Mixed forest</td>
<td>8.4</td>
<td>12.0</td>
<td>4.2</td>
<td>7.7</td>
<td>39.2</td>
<td>30.7</td>
<td>22.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Woodland/savanna</td>
<td>26.2</td>
<td>20.6</td>
<td>15.3</td>
<td>40.9</td>
<td>0.4</td>
<td>0.1</td>
<td>0.04</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tundra</td>
<td>18.3</td>
<td>20.3</td>
<td>10.2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1Within land cover datasets, mixtures of grasses and trees were categorized as woodland/savanna, and sparse vegetation or mixtures of grasses and shrubs were classified as tundra. Mixed needleleaf forests were classified as evergreen needleleaf (EN) trees. DB = deciduous broadleaf.

2Historical offline vegetation simulations

3CLM input PFTs (Lawrence and Chase, 2007)

4MCD12Q1 year 2003 (Friedl et al., 2002)

5Bontemps et al. (2010)

6North American version (Bartholome and Belward, 2005)

71km resolution (DeFries and Hansen, 2010)

85 minute resolution (Ramankutty and Foley, 2010)

9Global Land Cover Database Version 2.0, IGBP classification (Loveland et al., 2000)

10Global Land Cover Database Version 2.0, USGS classification (Loveland et al., 2000)
Table 2. Domain-wide annual surface energy budget means and anomalies for offline and online simulations.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Simulation</th>
<th>Historical (BAx1)</th>
<th>Changes(^1)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>BAx0</td>
</tr>
<tr>
<td>downwelling</td>
<td>offline</td>
<td>126.7</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>online</td>
<td>124.3 (0.34)</td>
<td>-0.74 (0.47)</td>
</tr>
<tr>
<td>shortwave</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>albedo</td>
<td>offline</td>
<td>0.266</td>
<td>-0.018</td>
</tr>
<tr>
<td></td>
<td>online</td>
<td>+0.249 (1.0x10(^{-3}))</td>
<td>-0.017 (1.3x10(^{-3}))</td>
</tr>
<tr>
<td>absorbed</td>
<td>offline</td>
<td>102.8</td>
<td>+1.57</td>
</tr>
<tr>
<td>shortwave</td>
<td>online</td>
<td>100.8</td>
<td>+1.00 (0.29)</td>
</tr>
<tr>
<td>downwelling</td>
<td>offline</td>
<td>249.1</td>
<td>0</td>
</tr>
<tr>
<td>longwave</td>
<td>online</td>
<td>266.6 (0.52)</td>
<td>+0.63 (0.73)</td>
</tr>
<tr>
<td>upwelling</td>
<td>offline</td>
<td>305.4</td>
<td>+0.12</td>
</tr>
<tr>
<td>longwave</td>
<td>online</td>
<td>314.8 (0.45)</td>
<td>+0.52 (0.64)</td>
</tr>
<tr>
<td>net radiation(^3)</td>
<td>offline</td>
<td>46.5</td>
<td>+1.45</td>
</tr>
<tr>
<td></td>
<td>online</td>
<td>52.5 (0.16)</td>
<td>+1.10 (0.21)</td>
</tr>
<tr>
<td>sensible heat(^3)</td>
<td>offline</td>
<td>29.1</td>
<td>+1.74</td>
</tr>
<tr>
<td></td>
<td>online</td>
<td>27.0 (0.13)</td>
<td>+0.79 (0.18)</td>
</tr>
<tr>
<td>latent heat(^3)</td>
<td>offline</td>
<td>15.2</td>
<td>-0.27</td>
</tr>
<tr>
<td></td>
<td>online</td>
<td>23.3 (0.08)</td>
<td>-0.31 (0.11)</td>
</tr>
</tbody>
</table>

\(^1\) Differences relative to BAx1. Values in parentheses represent 95% confidence interval widths for the differences.

\(^2\) Except albedo, units are W m\(^{-2}\)

\(^3\) Annual means, values in parentheses represent 95% confidence interval widths
Table 3. Domain-wide climate means and anomalies from coupled simulations.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Season</th>
<th>Historical</th>
<th>Changes</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>BAx0</td>
<td>BAx2</td>
</tr>
<tr>
<td>2-meter air temperature (°C)</td>
<td>annual</td>
<td>-0.1</td>
<td>+0.14 (0.16)</td>
</tr>
<tr>
<td></td>
<td>winter</td>
<td>-14.7</td>
<td>+0.08 (0.44)</td>
</tr>
<tr>
<td></td>
<td>spring</td>
<td>0.1</td>
<td>+0.31 (0.22)</td>
</tr>
<tr>
<td></td>
<td>summer</td>
<td>13.4</td>
<td>+0.13 (0.07)</td>
</tr>
<tr>
<td></td>
<td>fall</td>
<td>0.6</td>
<td>+0.03 (0.18)</td>
</tr>
<tr>
<td>precipitation (cm month⁻¹)</td>
<td>annual</td>
<td>5.74</td>
<td>+0.10 (0.05)</td>
</tr>
<tr>
<td></td>
<td>winter</td>
<td>4.93</td>
<td>+0.13 (0.11)</td>
</tr>
<tr>
<td></td>
<td>spring</td>
<td>4.60</td>
<td>+0.11 (0.11)</td>
</tr>
<tr>
<td></td>
<td>summer</td>
<td>6.63</td>
<td>+0.07 (0.08)</td>
</tr>
<tr>
<td></td>
<td>fall</td>
<td>6.79</td>
<td>+0.09 (0.13)</td>
</tr>
<tr>
<td>2-meter specific humidity (mg kg⁻¹)</td>
<td>annual</td>
<td>3.71</td>
<td>+0.02 (0.02)</td>
</tr>
<tr>
<td></td>
<td>winter</td>
<td>1.48</td>
<td>+0.02 (0.04)</td>
</tr>
<tr>
<td></td>
<td>spring</td>
<td>3.06</td>
<td>+0.05 (0.03)</td>
</tr>
<tr>
<td></td>
<td>summer</td>
<td>6.58</td>
<td>+0.01 (0.04)</td>
</tr>
<tr>
<td></td>
<td>fall</td>
<td>3.66</td>
<td>+0.003 (0.05)</td>
</tr>
<tr>
<td>vapor pressure deficit (Pa)³</td>
<td>annual</td>
<td>312.6</td>
<td>+3.5 (3.5)</td>
</tr>
<tr>
<td></td>
<td>winter</td>
<td>19.2</td>
<td>+0.7 (0.7)</td>
</tr>
<tr>
<td></td>
<td>spring</td>
<td>315.0</td>
<td>-0.9 (6.7)</td>
</tr>
<tr>
<td></td>
<td>summer</td>
<td>704.5</td>
<td>+13.3 (9.8)</td>
</tr>
<tr>
<td></td>
<td>fall</td>
<td>204.1</td>
<td>+0.8 (4.4)</td>
</tr>
<tr>
<td>boundary layer height (m)⁴</td>
<td>annual</td>
<td>403.5</td>
<td>+4.0 (1.6)</td>
</tr>
<tr>
<td></td>
<td>winter</td>
<td>213.6</td>
<td>+4.1 (2.4)</td>
</tr>
<tr>
<td></td>
<td>spring</td>
<td>463.7</td>
<td>+4.8 (3.4)</td>
</tr>
<tr>
<td></td>
<td>summer</td>
<td>586.5</td>
<td>+5.0 (3.2)</td>
</tr>
<tr>
<td></td>
<td>fall</td>
<td>345.2</td>
<td>+2.2 (2.2)</td>
</tr>
</tbody>
</table>

¹Differences relative to BAx1. Values in parentheses represent 95% confidence interval widths for the differences.
²Means for the corresponding time period.
³Calculated using mean relative humidity and mean daily maximum temperature
⁴Daily mean (including night)
Table 4. Differences in energy annual energy budgets between BAx2 and BAx1 for the main simulations and three sensitivity analyses

<table>
<thead>
<tr>
<th>Variable</th>
<th>Main</th>
<th>Sensitivity Analyses</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Add BA&lt;sup&gt;2&lt;/sup&gt;</td>
</tr>
<tr>
<td>albedo</td>
<td>+0.0132</td>
<td>+0.0163 (+23)</td>
</tr>
<tr>
<td>net radiation&lt;sup&gt;5&lt;/sup&gt;</td>
<td>-1.04</td>
<td>-1.39 (+34)</td>
</tr>
<tr>
<td>sensible heat&lt;sup&gt;6&lt;/sup&gt;</td>
<td>-1.23</td>
<td>-1.62 (+32)</td>
</tr>
<tr>
<td>latent heat&lt;sup&gt;6&lt;/sup&gt;</td>
<td>+0.172</td>
<td>+0.205 (+19)</td>
</tr>
</tbody>
</table>

<sup>1</sup>Values in parentheses represent percent changes from the main simulation differences.
<sup>2</sup>Burn area was added evenly to the boreal forest fraction of all grid cells, instead of in proportion to historical burn area, as was done in the main simulations.
<sup>3</sup>The probability of fire as a function of stand age was forced to pass through the point (0,0).
<sup>4</sup>The original deciduous broadleaf tree fraction during succession was halved, and the remainder given to evergreen needleleaf trees.
<sup>5</sup>W m<sup>-2</sup>

Table 5. Relationships between burn area and temperature

<table>
<thead>
<tr>
<th>Burn Area Unit</th>
<th>Season</th>
<th>Slope (°C per unit burn area)&lt;sup&gt;5&lt;/sup&gt;</th>
<th>Coupled Model Variance&lt;sup&gt;2&lt;/sup&gt;</th>
<th>Sensitivity Analysis Variance&lt;sup&gt;5&lt;/sup&gt;</th>
<th>Pooled Variance&lt;sup&gt;3&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>historical burn area</td>
<td>annual</td>
<td>-0.121</td>
<td>0.039</td>
<td>0.030</td>
<td>0.050 (41%)</td>
</tr>
<tr>
<td></td>
<td>winter - spring</td>
<td>-0.235</td>
<td>0.066</td>
<td>0.058</td>
<td>0.088 (37%)</td>
</tr>
<tr>
<td></td>
<td>February - April</td>
<td>-0.432</td>
<td>0.082</td>
<td>0.097</td>
<td>0.127 (29%)</td>
</tr>
<tr>
<td>Mha yr&lt;sup&gt;-1&lt;/sup&gt;</td>
<td>annual</td>
<td>-0.055</td>
<td>0.018</td>
<td>0.014</td>
<td>0.023 (41%)</td>
</tr>
<tr>
<td></td>
<td>winter - spring</td>
<td>-0.108</td>
<td>0.030</td>
<td>0.027</td>
<td>0.040 (37%)</td>
</tr>
<tr>
<td></td>
<td>February - April</td>
<td>-0.198</td>
<td>0.037</td>
<td>0.045</td>
<td>0.058 (29%)</td>
</tr>
</tbody>
</table>

<sup>1</sup>Derived from BAx0, BAx1, and BAx2 coupled simulations
<sup>2</sup>Calculated from 1x10<sup>6</sup> Monte Carlo simulations of slopes using coupled model standard errors of the means
<sup>3</sup>Variance from four sensitivity analyses: net radiation from three land model sensitivity analyses, and burn area reduction from the biological feedback analysis.
Figure 1. 500-meter land cover classification (a) and historical mean fire return intervals (FRIs) (b) within CLM grid cells in North American boreal domain.

Figure 2. Post-fire trajectories of burn probability (a), tree cover (b), evergreen needleleaf (EN) trees (b), shrubs (c), and generalized vegetation succession (d). In (a), a regression line was fit to annual probabilities of re-burn for years 0 - 60 from the Canadian National Fire Database and the Alaskan Large Fire Databases (NFDBs). The 'PF zero' line was forced through (0,0) and used for sensitivity analyses. (b – d) were developed using MODIS vegetation continuous fields (MOD44B) and land cover (MCD12Q1) over fire scars from the NFDBs. The shaded area in (b) represents ±1 standard error. A logistic curve was fit to EN trees, and a Weibull curve to shrubs. DB = ‘deciduous broadleaf’. The char type in (e) is included as part of the non-vegetated fraction.

Figure 3. Comparison of simulated to observed post-fire summer albedo (a), latent heat (b), and Bowen ratio (c), winter - spring albedo (d – e), and monthly albedo (f). Observations in (a) – (d) are extracted from Amiro et al. (2006) excluding the Southern Old Aspen site, (e) from Lyons et al. (2008), and (f) from Randerson et al. (2006). Modeled variables from July were chosen for (a) – (c) (observations spanned June 26 – July 24), January and February for (d) (observations spanned January 1 – March 1), and February and March for (e) (observations spanned February 2 – April 6). Modeled grid cells include field sites from Alaska (AK, 63.5°N, 145°W), Saskatchewan (SK, 54°N, 105°W), and Manitoba (MB, 55.9°N, 97.5°W). The young site in (f) includes years 3 – 5 post-fire, and mature includes approximate years 82 – 84. Only broadband albedo is considered. LE = latent heat, H = sensible heat, and \( R_n \) = net radiation.

Figure 4. Simulated plant functional types (PFTs) for the boreal forest fraction of an example grid cell with 300 sub-grid cells (a), and domain-wide PFT compositions under varying fire regimes (b). (b) includes vegetation from tundra pixels. Not shown are contributions from temperate vegetation, glaciers, wetlands, lakes, and urban areas.
Figure 5. Mean monthly historical (BAx1) and changes to energy budgets for offline and online simulations. Historical values are on left-hand axes, and anomalies on right-hand axes.

Figure 6. Mean monthly historical (BAx1) and changes to 2-meter air temperature. The shaded area in (a) delineates monthly standard error, and shaded regions in (b – d) represent monthly 95% confidence intervals.

Figure 7. Annual and February – April temperature anomalies across high-latitude North America, Greenland, and the surrounding ocean. Stipples represent 95% statistical significance.

Figure 8. Surface shortwave forcing (a) and 2-meter air temperature (b) changes across the domain as a function of burn area, for three time periods: annual, winter – spring, and February – April. Ellipse heights correspond to standard errors of the means. Net radiation and top of atmosphere energy balance (not shown) displayed similar patterns to (a).

Figure 9. Mean historical and changes to annual Northern Hemisphere sea ice coverage. Stipples represent 95% statistical significance.
burn area multiplier
-1.5
-1.0
-0.5
0.0
0.5

\[ \Delta oC \]

\[ \Delta (W \, m^{-2}) \]

a.

b.

annual
winter - spring
Feb - Apr