

# Effects of fire on regional evapotranspiration in the central Canadian boreal forest

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## Abstract

Changes in fire regimes are driving the carbon balance of much of the North American boreal forest, but few studies have examined fire-driven changes in evapotranspiration (ET) at a regional scale. This study used a version of the Biome-BGC process model with dynamic and competing vegetation types, and explicit spatial representation of a large ( $10^6$  km<sup>2</sup>) region, to simulate the effects of wildfire on ET and its components from 1948 to 2005 by comparing the fire dynamics of the 1948–1967 period with those of 1968–2005. Simulated ET averaged, over the entire temporal and spatial modeling domain,  $323 \text{ mm yr}^{-1}$ ; simulation results indicated that changes in fire in recent decades decreased regional ET by 1.4% over the entire simulation, and by 3.9% in the last 10 years (1996–2005). Conifers dominated the transpiration ( $E_C$ ) flux ( $120 \text{ mm yr}^{-1}$ ) but decreased by 18% relative to deciduous broadleaf trees in the last part of the 20th century, when increased fire resulted in increased soil evaporation, lower canopy evaporation, lower  $E_C$ , and a younger and more deciduous forest. Well- and poorly drained areas had similar rates of evaporation from the canopy and soil, but  $E_C$  was twice as high in the well-drained areas. Mosses comprised a significant part of the evaporative flux to the atmosphere ( $22 \text{ mm yr}^{-1}$ ). Modeled annual ET was correlated with net primary production, but not with temperature or precipitation; ET and its components were consistent with previous field and modeling studies. Wildfire is driving significant changes in hydrological processes by affecting mean stand age, forest species, and energy balance. These changes, particularly in poorly drained areas, may control the future carbon balance of the boreal forest.

*Keywords:* black spruce, boreal forest, disturbance, evapotranspiration, modeling, stand age, wildfire

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## Introduction

Significant changes in disturbance regimes have occurred for decades in the global boreal forest (Flannigan & Van Wagner, 1991; Kasischke & Turetsky, 2006), and future changes in boreal climate, CO<sub>2</sub>, and fire are likely to be large (Flannigan *et al.*, 2005). Given the large area and high carbon (C) storage of the global boreal forest (Flannigan & Van Wagner, 1991; Apps *et al.*, 1993; Gower *et al.*, 2001), understanding and predicting its responses to these factors is of great importance. A crucial part of this response will depend on how fire

disturbance and the hydrological cycle interact, particularly because so much of the boreal forest is poorly drained (National Wetlands Working Group, 1988). These areas sequester large amounts of C in peat, and the degree of their vulnerability to changes in climate, hydrology, and fire will influence regional biogeochemical cycles (Camill *et al.*, 2001).

Wildfire is the primary disturbance agent in most of the boreal forest (Stocks, 1991), and an important driver of ecosystem dynamics at both local and regional scales (Chambers & Chapin, 2002). The area burned by wildfire in central Canada has increased in recent decades (Stocks *et al.*, 2003), and modeling studies suggest that late 20th-century C balance was driven primarily by changes in the fire regime (Balshi *et al.*,

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2007; Bond-Lamberty *et al.*, 2007b). In central Canada, late 20th-century increases in fire has meant the conversion of large areas of mature evergreen needleleaf conifers (primarily *Picea mariana*) to stands dominated by young deciduous broadleaves (primarily *Populus tremuloides*), with significant consequences for regional C, nutrient, and water dynamics (Bond-Lamberty *et al.*, 2007b).

The effects of changing climate and disturbance on evapotranspiration (ET) and forest water cycling are not well understood. Human influences are changing ET in the northern hemisphere (Fernandes *et al.*, 2007), while at a smaller scale fire clearly affects stand water cycling (Ewers *et al.*, 2005; Amiro *et al.*, 2006a), watershed-scale hydrology (Valeo *et al.*, 2003), and regional climate (Chambers & Chapin, 2002). In the land-atmosphere water flux, ET is closely coupled with net primary production (NPP) through plant stomatal controls on transpiration (Campbell & Norman, 1998; Tang *et al.*, 2006); if ET changes with disturbance, this will have important implications for local, regional, and continental growth and climate dynamics. Increased fire and CO<sub>2</sub> are correlated (Flannigan *et al.*, 2005), and rising CO<sub>2</sub> may reduce ET and thus latent heat flux from the landscape, exacerbating regional climate change (Sellers *et al.*, 1996; Gedney *et al.*, 2006; Betts *et al.*, 2007).

Low temperatures and nitrogen availability constrain ET in the boreal forest (Baldocchi *et al.*, 2000; Ewers *et al.*, 2001), and boreal water vapor fluxes at the stand level vary with species composition (Linder & Troeng, 1980; Baldocchi *et al.*, 1997; Jarvis *et al.*, 1997), structural changes with stand age (Ewers *et al.*, 2005; Amiro *et al.*, 2006a, b), permafrost changes (Camill *et al.*, 2001), and soil drainage (Lafleur *et al.*, 1997) among other factors. The boreal tree canopy is dominated by relatively few species, predominantly evergreen needleleaf conifers, but mosses form a significant part of the C, nitrogen, and water cycles in many boreal forests (Skre & Oechel, 1981; Longton, 1992; Heijmans *et al.*, 2004; Suzuki *et al.*, 2007). The water dynamics of such bryophytes are not as well quantified as those of vascular plants, and are rarely modeled (Sonnentag *et al.*, 2008).

The goals of this study were to use Biome-BGC, a well-known ecophysiological model, to simulate the effects of changing wildfire on a subset of hydrological processes in the central Canadian boreal forest. Specifically, we examined the following questions: (i) what was the effect of changing fire on ET and its components in the last half-century? (ii) Given that the model predicts changes in dominance among the three vegetation types – coniferous trees, deciduous broadleaf trees, and mosses – that were simulated over this time period (Bond-Lamberty *et al.*, 2007b), what are the consequences for regional ET fluxes?

## Material and methods

### Simulation area

The region simulated was the BOREAS study region (Sellers *et al.*, 1995), a 1000 km × 1000 km bounded by the points (60.0°N, −111.0°W), (58.8°N, −93.5°W), (50.1°N, −97.0°W), and (51.0°N, −111.0°W). At 100 million ha, and comprising 6–8% of the global boreal forest and 15–20% of the North American boreal forest, it encompasses much of Manitoba and Saskatchewan as well as eastern Alberta, with Lake Winnipeg in the southeast, Hudson Bay in the northeast, and Lake Athabasca in the northwest (Fig. 1). Terrain is flat to gently rolling, with elevations ~ 250 m in the northeast and ~ 500 m in the southwest; species diversity is low and the forest is dominated by black spruce [*P. mariana* (Mill.) BSP], trembling aspen (*P. tremuloides* Michx.), and jack pine (*Pinus banksiana* Lamb.), with almost continuous bryophyte coverage (feathermosses and *Sphagnum* spp.) in the understory. From aspen parkland and mixedwoods in its south to black spruce-dominated boreal forest in the center to the forest-tundra transition in its north, this grid encompassed a wide cross-section of the boreal forest biome.

### The Biome-BGC model

The processes and logic of Biome-BGC have been discussed in detail in many publications (Running & Coughlan, 1988; Running & Gower, 1991; Kimball *et al.*, 1997a; Bond-Lamberty *et al.*, 2005a) and we present only a short summary here. Biome-BGC runs on a daily time step, using daily meteorological data, ecophysiological parameters, and general stand soil information to simulate energy, C, water, and nitrogen cycling (Running & Hunt, 1993; White *et al.*, 2000). Ecophysiological parameters define vegetation types – descriptions based on leaf habit, photosynthesis pathway, plant type, etc. – that are intended to be broadly significant. The version of Biome-BGC used here was based on the 4.1.2 version but extensively modified. These changes allowed Biome-BGC to more accurately simulate boreal soil temperatures (Bond-Lamberty *et al.*, 2005b), account for the deleterious effects of poor soil drainage on soil decomposition and vascular plant growth, and simulate bryophyte growth (Bond-Lamberty *et al.*, 2007a). This version also simulated multiple vegetation types concurrently, allowing for forest succession (Bond-Lamberty *et al.*, 2005a).

There are two ways in which water can enter a Biome-BGC simulation in our revised version of the model, via soil inflow and precipitation, and six ways in which it can exit: soil outflow, soil evaporation, snow

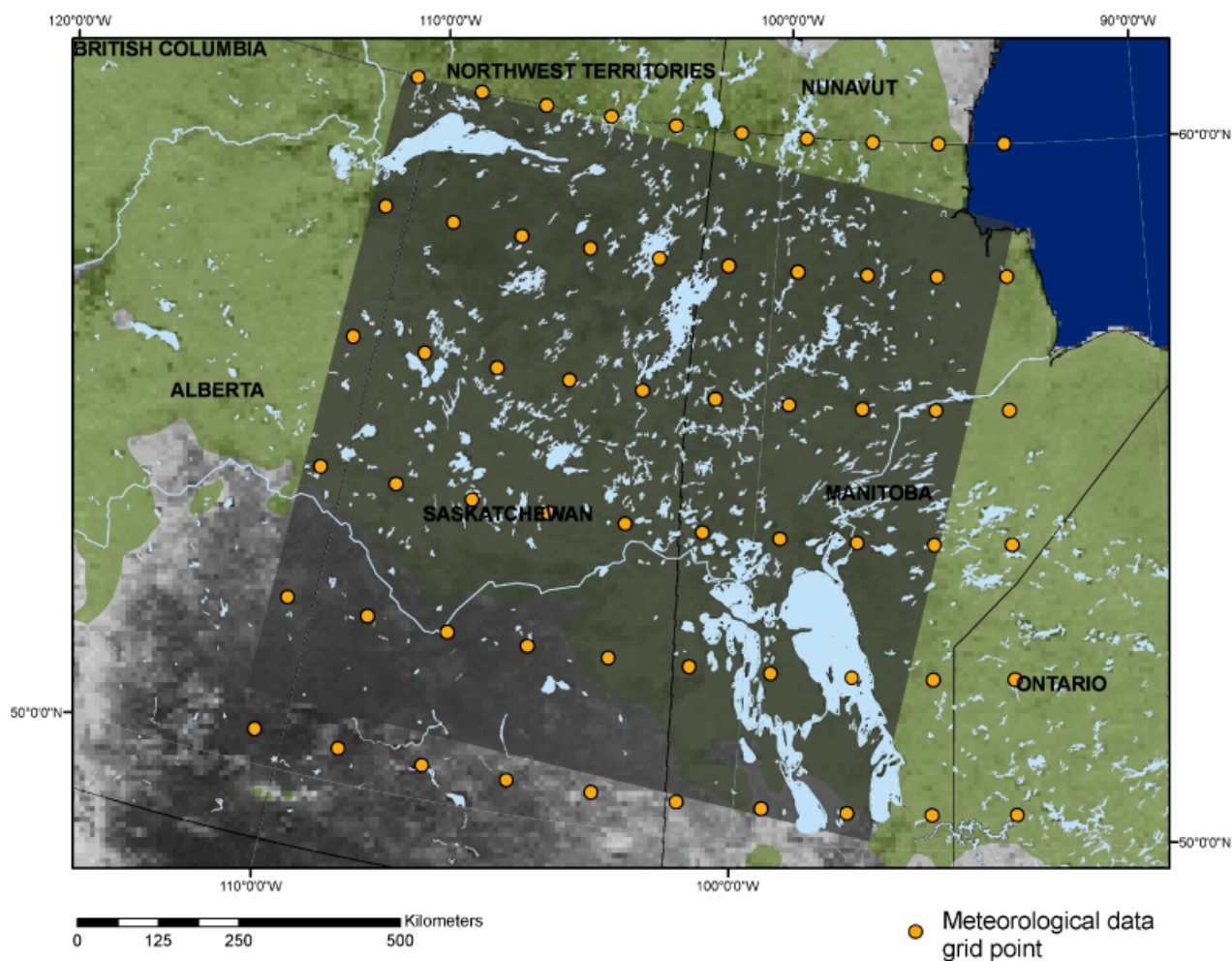


Fig. 1 The BOREAS grid (1000 km × 1000 km) simulated in this study, with yellow dots showing NCAR meteorological data points.

sublimation, canopy evaporation, canopy transpiration, and volatilization in fire. A fraction of precipitation is intercepted by the plant canopy and subject to evaporation before transpiration occurs. Soil evaporation and canopy ET are determined using the Penman–Monteith equation (Monteith, 1965; Campbell & Norman, 1998) and thus driven primarily by radiation and vapor pressure deficit. Leaf and canopy conductances are determined by stomatal, cuticular and boundary layer conductances (the first two in parallel, the last in series), defined on a vegetation-specific basis and modulated by plant water stress, temperature, vapor pressure deficit and soil anoxia (Thornton, 1998; Bond-Lamberty *et al.*, 2007a). All vegetation types being simulated have the potential to contribute to total transpiration, although in the current study moss conductance was set to zero (see the following text) and thus its effect was only via canopy evaporation and shading of the soil surface.

#### *Sources of input data: land cover and soils*

Forested vs. nonforested land cover categories were determined using the IGBP landcover classification of the MODIS landcover product for 2004 (MOD12Q1, <http://www-modis.bu.edu/landcover/userguidelc/lc.html>). If the native projection of these (and any other) data was not the BOREAS Albers Equal Area Conic (BOREAS AEAC), then the data were reprojected. MODIS products were imported, subset, reprojected to the AEAC, and masked to include only the simulation grid; nearest neighbor resampling was used for land cover.

Soil data were derived from data layers of the BOREAS regional soils data in raster format ([http://daac.ornl.gov/BOREAS/guides/Region\\_Soils\\_Raster.html](http://daac.ornl.gov/BOREAS/guides/Region_Soils_Raster.html)). Estimates of the sand, silt, and clay percentages were obtained using the soil texture raster data and the Canadian Soil Information System (CanSIS) website

**Table 1** Summary of meteorological forcing data, by grid quadrant (northwest, northeast, southwest, southeast) for years 1948–2005

Quadrant	Precipitation		$T_{\text{AVG}}$ (°C)		$T_{\text{MIN}}$ (°C)		$T_{\text{MAX}}$ (°C)	
	Mean*	Trend	Mean	Trend	Mean†	Trend	Mean	Trend
NW	0.20	<b>-0.01</b>	-1.21	0.05	-7.39	0.10	1.94	0.03
NE	0.17	<b>-0.01</b>	-2.95	0.03	-9.00	0.10	0.21	0.00
SW	0.17	0.00	3.20	-0.07	-3.60	0.05	6.68	-0.13
SE	0.19	<b>0.00</b>	2.35	-0.06	-4.19	0.04	5.76	-0.12
All	0.18	<b>0.00</b>	0.35	-0.01	-6.05	0.07	3.65	-0.06

Mean ( $N = 100$  met data points; cf. Fig. 1) and trend (per 10 years) data are shown for precipitation ( $\text{cm day}^{-1}$ ), and mean annual daily mean, minimum, and maximum temperatures ( $T_{\text{AVG}}$ ,  $T_{\text{MIN}}$ , and  $T_{\text{MAX}}$ , respectively; °C). Significant trends ( $P < 0.05$  in a simple linear model) are in bold.

\*Standard errors (based on annual means) were 0.02–0.03 mm, 0.91–1.02 °C, 1.10–1.21 °C, and 0.98–1.09 °C, respectively.

†For 1948–2005, mean annual number of days with  $T_{\text{MIN}} > 0$  °C was 143, 148, 231, and 228 for NW, NE, SW, and SE quadrants, respectively; trends (per 10 years) were 0.1, 1.0, 0.6, and 1.0.

(<http://sis.agr.gc.ca/cansis/>); these were applied to all other soils in the raster image that were similar to known soil textures. Thirty percent of the BOREAS grid was coded as ‘not applicable’. To increase the number of pixels with soil texture information, these ‘not applicable’ locations were filled in by association with land cover. The rooting depth data layer was used to calculate the Biome-BGC effective soil depth parameter, based on four categories (0.2, 0.75, 1.5, and 2.5 m).

A digital elevation model, obtained from the ORNL DAAC, had a spatial resolution of 1 km and elevations estimated to the nearest meter. The MODIS MCD43B product (<http://www-modis.bu.edu/brdf/userguide/param.html>) was used for albedo. A 16-day composite image from July 28 to August 12, 2003, was calibrated and resampled using bilinear interpolation, and the shortwave band extracted from the 20-band image. A soil drainage class raster image was used to compute water inflow and outflow rate parameters (inflow  $\alpha_0$ , surface outflow  $\alpha_1$ , and subsurface outflow  $\alpha_2$ ) (Bond-Lamberty *et al.*, 2007a) based on the following criteria: excessively well drained ( $\alpha_0 = 0.0$ ,  $\alpha_1 = 1.0$ ,  $\alpha_2 = 0.8$ ); well to moderately well drained ( $\alpha_0 = 0.0$ ,  $\alpha_1 = 1.0$ ,  $\alpha_2 = 0.5$ ); poorly drained ( $\alpha_0 = 0.5$ ,  $\alpha_1 = 0.1$ ,  $\alpha_2 = 0.01$ ); very poorly drained ( $\alpha_0 = 1.0$ ,  $\alpha_1 = 0.01$ ,  $\alpha_2 = 0.0$ ).

#### Sources of input data: meteorology and fire

Meteorological data were downloaded from the Climate Analysis Branch at NOAA (<http://www.cdc.noaa.gov/cdc/reanalysis/reanalysis.shtml>) for 1948–2005. These data included maximum daily temperature, minimum daily temperature, specific humidity, shortwave radiation, surface pressure, and precipitation rate, and were on the T1 Gaussian grid. A subset of the global grid was

chosen to adequately cover the study area, and the data resampled to the simulation grid, with 20 km resolution using weighted distance interpolation with the six nearest latitude/longitude points (Fig. 1). Vapor pressure deficit was calculated from surface pressure and specific humidity (Campbell & Norman, 1998). Meteorology input data are summarized in Table 1.

Spatially and temporally explicit data from the Canadian Forest Service’s Large Fire Database (LFDB) were used to drive the model; these data include 97% of the area burned in Canada (Stocks *et al.*, 2003). The LFDB covers 1959–2005; data for 1948–1958 were derived based on provincial records of total area burned and the known fire size distributions data of the LFDB. Mean area burned in the study area was  $122\,773 \text{ ha yr}^{-1}$  from 1948 to 1967 and  $229\,891 \text{ ha yr}^{-1}$  from 1968 to 2005, an 87% increase. The Biome-BGC disturbance mechanism used here was developed by the Forest Ecosystem Ecology Laboratory at the University of Wisconsin–Madison. Fire events (of constant intensity and length of 1 day) occurred spatially following to the LFDB and on a random day of the growing season. In each fire event, 10%, 90%, 5%, 10%, 80%, 40%, and 10% of C was volatilized from stem, leaf, coarse root, fine root, litter, woody debris, and shallow soil, respectively (cf. Albin & Reinhardt, 1995; Arora & Boer, 2005; Neff *et al.*, 2005), with the balance going to woody debris (for stems and roots), litter (for leaves), or remaining unaffected (for soil). In addition, 90% of mineral soil nitrogen was volatilized. Plant growth was restarted the year following fire. Plant physiological parameters for each functional group were standardized across the entire simulation grid (i.e., did not vary spatially). We used the values given by Bond-Lamberty *et al.* (2006); a summary of these parameters is given in Table 2.

**Table 2** Summary of physiological parameters used in simulations, by vegetation type

Parameter	Spruce	Aspen	Bryophyte
Fine root:leaf C	1.6	1.5	0.05
Stem C:leaf C	2.3	2.0	n/a
Live:total wood C	0.1	0.1	n/a
Coarse root:stem C	0.3	0.4	n/a
Leaf C:N	59.6	19.8	40.0
Leaf litter C:N	90.0	34.6	87.0
Fine root C:N	60.0	19.8	40.0
SLA (projected area basis, $\text{m}^2\text{kg}^{-1}\text{C}$ )	8.3	38.7	20.0
All sided:projected leaf area	3.1	2.0	2.0
Leaf N in Rubisco (%)	6.0	14.0	6.0
Max stomatal conductance ( $g_s$ , $\text{mm s}^{-1}$ )	3.0	5.0	0.0
Cuticular conductance ( $\text{mm s}^{-1}$ )	0.01	0.01	0.03
Boundary layer conductance ( $\text{mm s}^{-1}$ )	80.0	20.0	50.0
$\Psi_L$ start of $g_s$ reduction (MPa)	-0.5	-0.5	n/a
$\Psi_L$ complete $g_s$ reduction (MPa)	-1.7	-2.3	n/a
VPD start of $g_s$ reduction (kPa)	1.0	1.0	n/a
VPD complete $g_s$ reduction (kPa)	4.1	4.0	n/a
Flooded $g_s$ reduction (%)	30	20	n/a
Days to reach $g_s$ reduction (days)	5	5	n/a
Full turgor water content ( $\text{H}_2\text{O}:\text{C}$ )	n/a	n/a	5.0
External water (%)	n/a	n/a	50

Full data and sources can be found in Bond-Lamberty *et al.* (2006).

### Simulation procedures

The simulation grid was  $10^6 \text{ km}^2$ ; simulation cell size was  $1 \text{ km}^2$ . Each grid cell was independently spun up under historical conditions, using the parameters in Tables 1 and 2 and with a fire return interval of 175 years, until just before its deep soil C stabilized (i.e., leaving a small ongoing C sink) (Pietsch & Hasenauer, 2006). From this point, two 58-year simulations were performed: one in which disturbance remained at 1948–1967 levels (scenario ‘MC’, i.e., meteorology plus climate), and one in which it followed historical records throughout the simulation (scenario ‘MDC’, i.e., meteorology plus climate plus disturbance). In both scenarios, climate and  $\text{CO}_2$  followed their historical records (cf. Bond-Lamberty *et al.*, 2007b). All simulations were performed on a distributed computing grid. Sixty desktop computers in a public computer laboratory at the University of Wisconsin–Madison constituted the grid’s core; Apple Computer’s Xgrid™ software (Apple Computer, Cupertino, CA, USA) handled job distribution and processing, while custom

shell and IDL (ITT Visual Information Solutions, Boulder, CO) scripts were used for pre- and postprocessing of data. Statistical analyses were conducted using R version 2.6.0 (R Development Core Team, 2007).

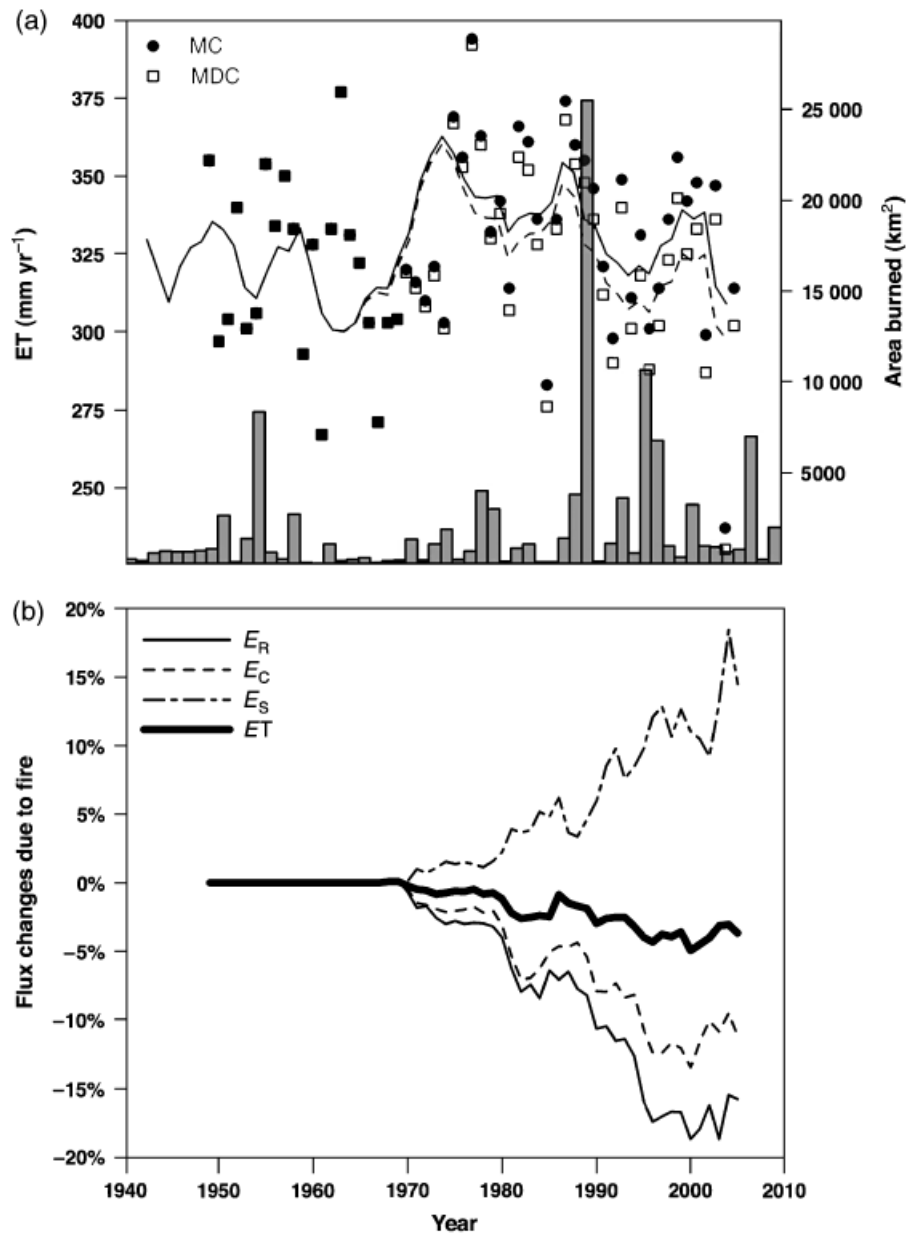
### Results

Increases in wildfire after 1970 had strong effects on ET and its components (Fig. 2). ET averaged  $323 \pm 30 \text{ mm yr}^{-1}$  over the entire simulation (the entire modeling domain in space and time; error term is interannual standard deviation), and was 1.4% lower in the real-world ‘MDC’ scenario relative to the ‘MC’ control scenario; over the last 10 years of the simulation, however, this difference was 3.9% (Fig. 2a). Increased wildfire resulted in increased soil evaporation (+13% from 1996 to 2005), as more sunlight reached the surface in newly burned stands, but lower canopy evaporation and transpiration (–17% and –12% over the same time period, respectively; Fig. 2b) as leaf area was reduced by fire.

Across the region, canopy transpiration ( $E_C$ ,  $140 \pm 11 \text{ mm yr}^{-1}$ ) and evaporation from the soil surface ( $E_S$ ,  $139 \pm 21 \text{ mm yr}^{-1}$ ) comprised the largest percentage of total ET (Table 3). Evaporation from the canopy ( $E_R$ ) comprised 14–17% of ET,  $44 \text{ mm yr}^{-1}$ , and decreased in the real-world MDC scenario as mature stands with high leaf area burned preferentially. Conifers dominated the  $E_C$  flux ( $120 \text{ mm yr}^{-1}$  in the real-world MDC scenario); transpiration from deciduous trees and bryophytes comprised similar percentages of ET. Fire, as noted above, sharply increased the evaporative flux from the soil surface to the atmosphere (Table 3). In contrast, conifers and bryophytes together dominated the  $E_R$  flux, the latter due to their ectohydric nature (ability to carry water external to the plant); each comprised 49% of the  $E_R$  flux ( $\sim 22 \text{ mm yr}^{-1}$ ) in both scenarios (data not shown).

The spatial distribution of changes in the ET flux between the low- (MC) and high-fire (MDC) scenarios is shown in Fig. 3. The depressing effects of fire on ET are clearly visible in the fire scars (circles) running from northwest to southeast across the BOREAS study region. Well-drained areas had higher ET than poorly drained areas in the model, due to higher  $E_C$  ( $56 \pm 5 \text{ mm yr}^{-1}$  vs.  $29 \pm 3 \text{ mm yr}^{-1}$ ),  $E_R$  ( $15 \pm 2 \text{ mm yr}^{-1}$ ), and  $E_S$  ( $139 \pm 21 \text{ mm day}^{-1}$ ) fluxes did not vary by soil drainage class.

The driver behind changes in  $E_C$  was the shift in species due to fire conversion of mature evergreen conifers to young broadleaf deciduous stands (Fig. 4). Transpiration by evergreen conifers, deciduous broadleaves, and mosses changed by –25, 9, and –2  $\text{mm yr}^{-1}$ , respectively, due to changing fire regimes. For 1995–



**Fig. 2** Effects of fire regimes changes on evapotranspiration (ET) and its components over the course of the simulation. (a) Mean daily ET under the mid-20th-century (MC) and actual historical (MDC) fire regimes; lines (solid line for MC, dashed for MDC) show the 5-year running mean. The bar graph (right-hand axis) shows area burned in the study area. (b) Change in ET and its components (canopy evaporation  $E_R$ , plant transpiration  $E_C$ , and soil evaporation  $E_S$ ) in the historical MDC scenario relative to the mid-20th-century MC scenario.

2005, for the final 10 years of the simulation, this represented  $E_C$  changes of  $-18\%$ ,  $148\%$ , and  $-18\%$  of the prefire values for each vegetation type, respectively (Table 3).

The components of modeled annual ET tended to be strongly correlated with temperature ( $T_{AVG}$ ), precipitation, and NPP (Fig. 5). Significant relationships in Fig. 5 were those between  $E_R$  and all three variables (for  $T_{AVG}$

and NPP,  $R^2 = 0.13$ ; for precipitation,  $F_{1,56} = 48.18$ ;  $P < 0.001$ ;  $R^2 = 0.46$ ); between  $E_C$  and  $T_{AVG}$  ( $F_{1,56} = 9.27$ ;  $P = 0.004$ ;  $R^2 = 0.14$ ); between  $E_S$  and precipitation ( $F_{1,56} = 4.62$ ;  $P = 0.036$ ;  $R^2 = 0.08$ ); and between total ET and NPP ( $F_{1,56} = 4.63$ ;  $P = 0.036$ ;  $R^2 = 0.15$ ; not shown). The large interannual variability of precipitation and forest growth meant that  $R^2$  values tended to be low, even for these significant correlations.

**Table 3** Components of total evapotranspiration (ET) as a percentage of the total ET flux, by simulation scenario (MC, mid-20th-century fire data and MDC, historical data)

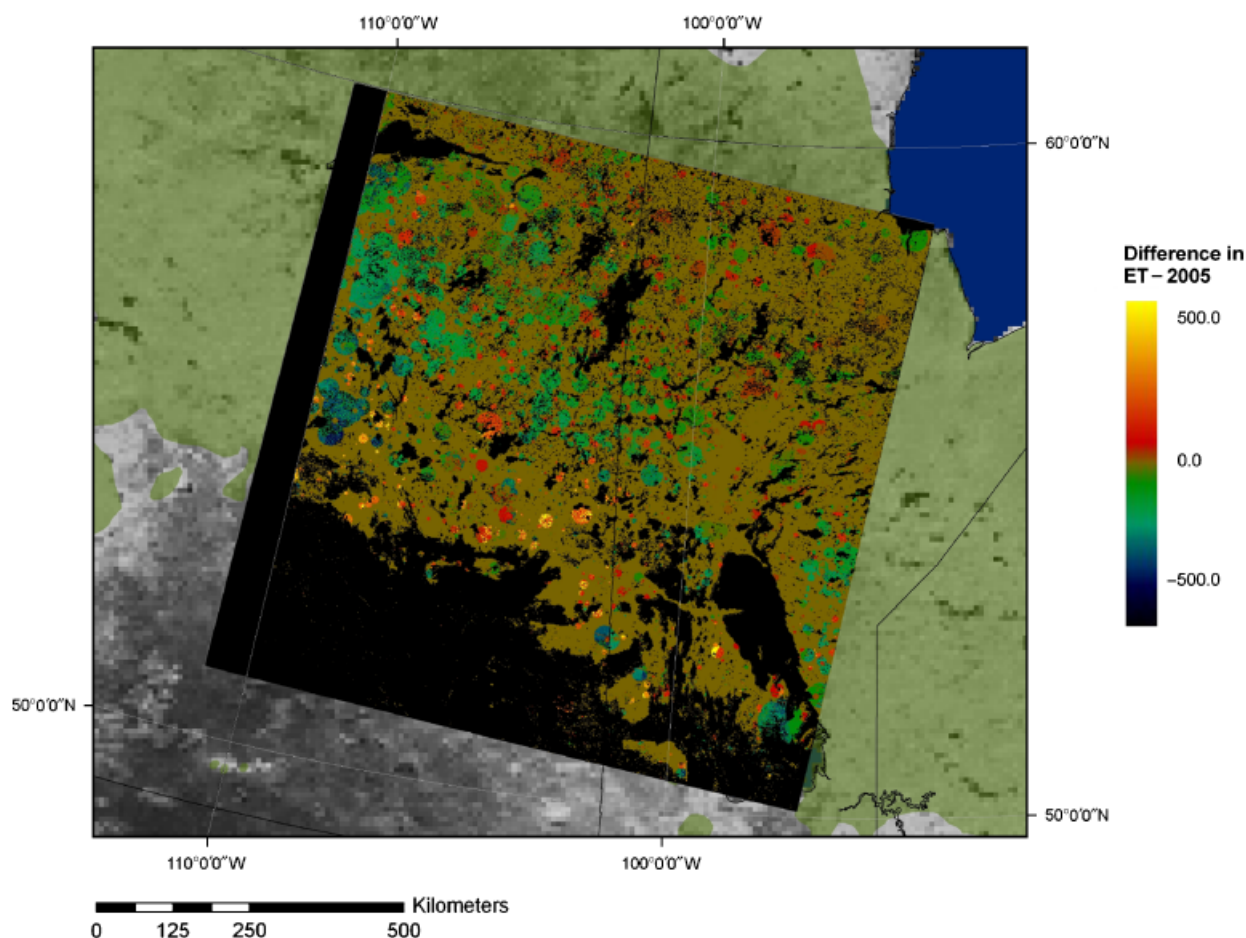
Scenario	$E_R$ (%)	$E_S$ (%)	$E_C$ percentage of ET (%)			
			Total	ENF	DBF	BRY
1968–2005						
MC	15	39	46	41	3	3
MDC	14	42	45	38	4	2
1996–2005						
MC	17	36	47	42	2	3
MDC	15	42	43	35	5	3

Canopy evaporation ( $E_R$ ), soil evaporation ( $E_S$ ), and canopy transpiration ( $E_C$ ) are shown;  $E_C$  is broken out by vegetation type (ENF, evergreen needleleaf; DBF, deciduous broadleaf; BRY, bryophytes).

## Discussion

### Comparison with field data: trees and mosses

Boreal forest ET has been measured at 1.5–2.0 mm day<sup>-1</sup> during the growing season (Lafleur, 1992; Grelle *et al.*, 1997; Jarvis *et al.*, 1997; Kelliher *et al.*, 1997; Amiro *et al.*, 2006a), implying  $\sim 250$  mm yr<sup>-1</sup> over the year if one assumes a 100-day shoulder season flux of 0.75 mm day<sup>-1</sup> and winter flux of 0.15 mm day<sup>-1</sup>. Arain *et al.* (2003) measured ET for 2 years in a black spruce forest in northern Saskatchewan, Canada, and reported a maximum ET of 3.5 mm day<sup>-1</sup>, winter fluxes of 0.1–0.25 mm day<sup>-1</sup>, and a 2-year mean of 356 mm yr<sup>-1</sup>, compared with the mean value of 323 mm yr<sup>-1</sup> reported here. A comparison of model results against a range of eddy covariance data from across western Canada is shown in Fig. 6; the mean and variability of ET simulated here broadly matches the mean and range of these published data.



**Fig. 3** Effect of changing fire regime on simulated evapotranspiration (ET) at the regional scale. Colors indicate difference in mean daily ET (mm yr<sup>-1</sup>) between the historical and mid-20th-century scenarios. Data are from 1996 to 2005 model output.

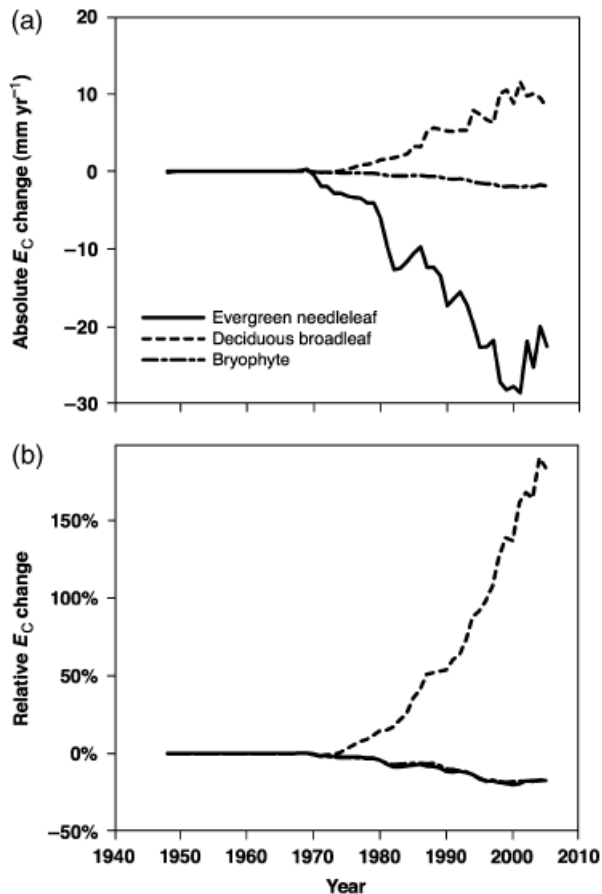


Fig. 4 Absolute (a) and relative (b) changes in canopy transpiration ( $E_C$ ) under actual fire regimes, relative to the mid-20th-century regime, by vegetation type and year.

In the results here,  $E_C$  was dominated by evergreen conifers, but the conifers' contribution was subject to large changes with stand age and fire. Ewers *et al.* (2005) used Granier sapflux sensors to calculate growing season  $E_C$  values of  $0.1\text{--}3.3\text{ mm day}^{-1}$  for a boreal black spruce chronosequence; broadleaf *P. tremuloides* accounted for from 33% of  $E_C$  (in the youngest, 12-year-old stand) to 0% (in the oldest stands) of  $E_C$ , implying a chronosequence average value of  $\sim 3\%$ . The results of this study are similar to those of Ewers *et al.* (2005), as broadleaf trees accounted for 4–5% of  $E_C$  at the regional scale here. Grelle *et al.* (1997) found that  $E_C$ ,  $E_R$ , and  $E_S$  comprised 65%, 20%, and 15% of total ET in a Swedish pine/spruce forest; the corresponding percentages in this simulation were 43%, 15%, and 42%, that is, soil evaporation comprised a larger percentage of ET, and canopy transpiration a small percentage, than in the 100-year-old stand measured by Grelle *et al.* (1997). This is reasonable, as our regional-level results included many recently burned stands with high  $E_S$ .

The effect of soil drainage in the model was expressed through differences in  $E_C$ , as neither  $E_R$  nor  $E_S$  varied by

drainage. The higher  $E_C$  and ET simulated here is consistent with energy balance data from a black spruce chronosequence in Manitoba (B. Amiro, University of Manitoba, unpublished data), but there have been few other studies of paired stands. Bogs and other poorly drained areas are dominated by mosses, which play an important role in cycling of water in boreal forests (Skre & Oechel, 1981; Longton, 1992; Price *et al.*, 1997; Heijmans *et al.*, 2004), but because their physiological controls on water use differ radically from those of vascular plants (Proctor, 2000), they have rarely been modeled as plants *per se* (Zhang *et al.*, 2002; Bond-Lamberty *et al.*, 2007a). Heijmans *et al.* (2004) reported moss evaporation rates of  $0.3\text{--}1.5\text{ mm day}^{-1}$  in an Alaskan boreal forest during the growing season, corresponding roughly to  $40\text{--}160\text{ mm yr}^{-1}$  on an annual basis (assuming shoulder season and winter fluxes of  $0.1$  and  $0.0\text{ mm day}^{-1}$ , respectively). Simulated bryophyte water flux at a regional level averaged  $22\text{ mm yr}^{-1}$  in this study; assuming poorly drained areas comprise about one-third of the landscape (National Wetlands Working Group, 1988), this results compares well with the results of Heijmans *et al.* (2004).

#### Comparison with large-scale modeling studies

Few large-scale modeling studies of the hydrologic cycle have been performed for the boreal forest. Kang *et al.* (2006) used Biome-BGC to model  $357\,500\text{ km}^2$  in the BOREAS region from 1959 to 1996, and examined the effects of changing  $\text{CO}_2$ , climate, and fire on NPP and ET. Their results are particularly appropriate to discuss here, as their study overlaps with ours in geographic area, time span studied, and model used. Kang *et al.* (2006) found that (i) ET was most sensitive to precipitation changes, and (ii) fire decreased ET by 9–38%, depending on forest type, while increasing soil evaporation. We found that ET was not significantly correlated with precipitation, in contrast to finding (i) by Kang *et al.* (2006), with the primary driver of this sensitivity being the correlations between  $E_R$  and  $E_S$  with precipitation; the correlation with transpiration was not significant, although precipitation was significantly correlated with both  $E_S$  and  $E_R$ . This is unsurprising as water is probably not a limiting factor for plant growth in these systems. Our finding that fire increases  $E_S$  while reducing overall ET was similar to that of Kang *et al.* (2006), though we observed much smaller decreases. This difference is probably due to the multi-vegetation model used here, in which a coniferous stand can be quickly replaced after burning by a deciduous stand. The drop in leaf area index (LAI) after fire is relatively transient, and the ET response a balance of decreased  $E_C$  and increased  $E_S$ ; such a response is



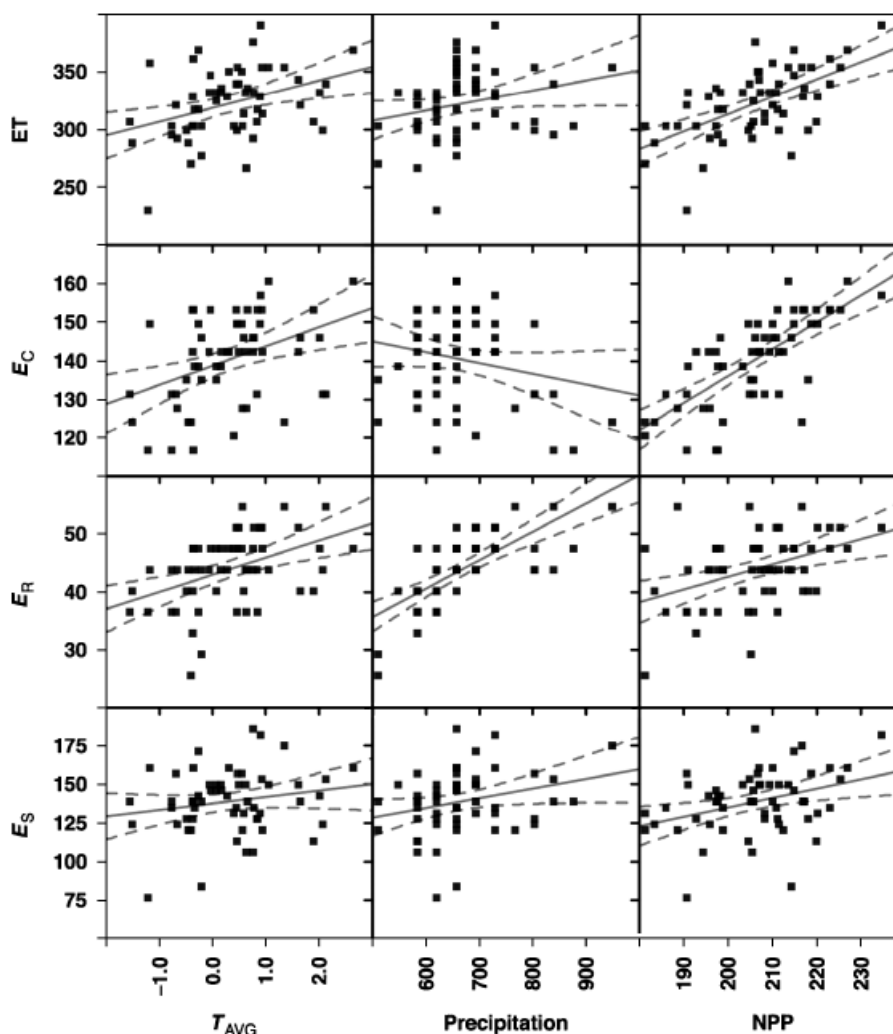


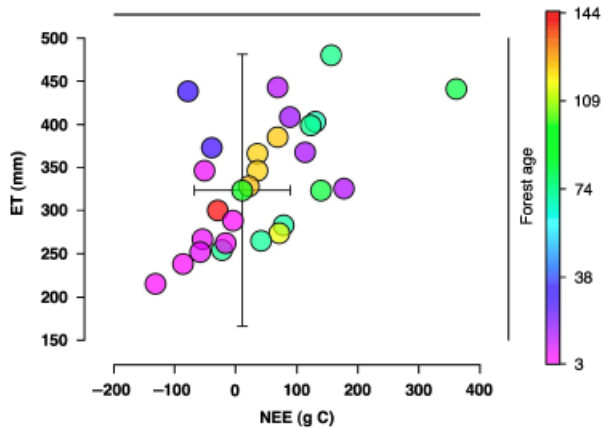
Fig. 5 Relationship of total evapotranspiration (ET), canopy evaporation ( $E_R$ ), transpiration ( $E_C$ ), and soil evaporation ( $E_S$ ) (all in  $\text{mm yr}^{-1}$ ) to mean annual temperature ( $T_{\text{AVG}}$ ,  $^{\circ}\text{C}$ ), precipitation ( $\text{mm yr}^{-1}$ ), and net primary production (NPP;  $\text{g C m}^{-2} \text{yr}^{-1}$ ). Dotted lines show 95% confidence intervals.

similar to arguments made for ET homeostasis (Roberts, 1983). In summary, our results are largely consistent with those of Kang *et al.* (2006).

#### *The ability of Biome-BGC to simulate water fluxes*

Numerous authors have examined the accuracy of the Biome-BGC in simulating boreal ET at a daily time step. In the well-drained BOREAS sites, Kimball *et al.* (1997b) reported that Biome-BGC explained 62–98% of the variability in observed daily ET and soil water, with black spruce in the model rarely subject to water stress. Amthor *et al.* (2001) found that Biome-BGC underpredicted monthly ET for the moderately well-drained BOREAS NSA tower site, ranking in the lower third of the nine models examined. On a daily time step, however, Biome-BGC performed in the top third

(Amthor *et al.*, 2001). Across seven temperate North American evergreen forests, the model had a mixed performance simulating annual ET, greatly underestimating it for a Florida *Pinus elliottii* plantation (Thornton *et al.*, 2002). Churkina *et al.* (2003) found that Biome-BGC explained 60–88% of measured ET variability in four European coniferous forests, while overestimating ET at high fluxes. Pietsch *et al.* (2003) reported that extending Biome-BGC to allow for groundwater infiltration and flooding – key characteristics of floodplains – greatly improved the model's performance. Similarly, Engstrom *et al.* (2006) modified the model by adding water storage and nonvascular vegetation evaporation routines, and found that while the original Biome-BGC greatly underestimated ET, their new version slightly overestimated it at well- and poorly drained Arctic tundra sites.



**Fig. 6** Annual evapotranspiration (ET) vs. net ecosystem exchange (NEE), simulated and observed data. Most circles show published data from eddy covariance sites in the western Canadian boreal forest (Black *et al.*, 1996; Jarvis *et al.*, 1997; Goulden *et al.*, 1998; Arain *et al.*, 2003; Griffis *et al.*, 2003; Amiro *et al.*, 2006a; Pejam *et al.*, 2006; Sass, 2007). The circle with error bars is the mean simulated ET and NEE over the modeling domain; error bars are spatial variability. Circle fill color indicates forest age.

There are weaknesses in the modeling approach used here. We did not consider the effects of grasses or shrubs, and their contribution to ET flux can be significant in some boreal systems (e.g., Lafleur *et al.*, 1997; Admiral *et al.*, 2006). Biome-BGC includes no explicit energy balance calculation, which simplifies the input parameter requirements but forces a crude calculation of, for example, soil temperatures (Bond-Lamberty *et al.*, 2005b). Finally, Biome-BGC is not designed to simulate poorly drained areas, and its performance in such areas is significantly worse than in well-drained forests (Bond-Lamberty *et al.*, 2006, 2007a). The drivers of ET are considerably different in poorly drained forests (Mackay *et al.*, 2007): we expect a wetland/peatland forest to be driven by net radiation in the early stages after fire and to become more (but never completely) driven by VPD in the latter stages of succession. This also implies that the drivers of ET will be evaporation from bryophytes and soil early in succession, and tree transpiration later in succession, and that interannual variability of ET will be lower in such areas.

#### *Effects of disturbance on forest water and C fluxes*

Boreal wildfires in central Canada effect species shifts over decadal periods, converting evergreen conifer (typically *P. mariana* or *P. banksiana*) stands to broadleaf deciduous ones (*P. tremuloides*), as well as changing the

age structure of the forest (Kurz & Apps, 1999). In addition, other disturbances such as logging, insects, and pathogens are growing in importance and can have effects comparable in scale to those from fire (Malmström & Raffa, 2000; Kurz *et al.*, 2008). The resulting change in species and leaf phenology will in turn affect the seasonality of C and water fluxes (Ewers *et al.*, 2005). Even without species changes, a shift in mean stand age may be important: Goulden *et al.* (1998) concluded that high evaporative demand ( $D$ ) would have little effect on the gross primary production of a mature black spruce forest, but younger black spruce stands have been shown to be more sensitive to  $D$  than older forest (Ewers *et al.*, 2005). With increasing fire, such older stands constitute a shrinking proportion of the forest and thus cannot be taken as representative of the region in which they sit.

These changes also affect the underlying soils. The removal of the dense canopy, and frequently the insulating organic layers, means increased soil evaporation, reduced transpiration and canopy evaporation, and concomitant shifts in the temperature regimes of the soil (Bond-Lamberty *et al.*, 2005b). Water and thick insulating moss and soil layers minimize decomposition in poorly drained boreal areas (Viereck, 1983; Harden *et al.*, 1997); this results in forests with the highest C density in the world, but also means that such C reservoirs are vulnerable to changes in fire and drainage regimes, although they may currently be damping swings in regional C balance (Bond-Lamberty *et al.*, 2007b). Changes in hydrological processes may thus control the future C balance of the boreal forest (Goulden *et al.*, 1998; Barr *et al.*, 2006; Dunn *et al.*, 2006; Krishnan *et al.*, 2006), and as such should be a focus of improvement in ecophysiological models.

#### **Conclusion**

Wildfire is driving significant changes in hydrological processes in the boreal forest, changes that in turn may affect the growth of the forest, its ability to sequester C, and regional climate. This makes understanding the complete forest disturbance history – in the central Canadian boreal forest, fire, insects, and logging – critical. In addition, the biotic and abiotic differences between well- and poorly drained areas mean that changing disturbance regimes will affect them quite differently. Ecosystem models that scale across large areas, accurately simulate poorly drained areas, and have been tested against robust ET data from forests of a variety of drainage and stand-age points, will be necessary for the accurate estimation and modeling of boreal hydrological dynamics.

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