Simulating the effects of past changes in climate, atmospheric composition, and fire disturbance on soil carbon in Canada's forests and wetlands

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Received 18 January 2007; revised 24 March 2008; accepted 29 April 2008; published 2 August 2008.

[1] Using the Integrated Terrestrial Ecosystem Carbon Cycle model (InTEC), six simulations with different input scenarios of climate, CO_2, and nitrogen (N) deposition are conducted to study the changes of soil carbon (C) content in Canada's forests and wetlands during 1901–2000. Simulated total C stored in Canada's forest and wetland soils is 164.5 Pg C and accounts for about 7% of the global total of 2400 Pg C to the depth of 2 m, implying the significance of Canada's forest and wetland soils in the global terrestrial C cycle. Soils of Canadian forests and wetlands sequestrated 3.9 Pg C (2.6 Pg C in forests and 1.3 Pg C in wetlands) during 1901–2000 because of the integrated effects of climate, CO_2 fertilization, N deposition, and forest age factors. The changes of soil C content during 1901–2000 ranged spatially from -2 kg C m^{-2} to 4 kg C m^{-2}, depending on fire disturbance history, climate change pattern, and N deposition rates. Soil C increased by 2 to 4 kg C m^{-2} in Eastern Hudson Plains, Eastern Middle Boreal Shield, Southern Boreal Shield, and Atlantic Maritime and decreased by more than 1 kg C m^{-2} in Southern Boreal Plains. Simulations shows that climate influences growing conditions, growing season length, net N mineralization, and N fixation and therefore was the biggest driver of the increase in total soil C content during 1901–2000, followed by CO_2 fertilization and N deposition. The climate-induced increase of soil C occurred mainly in the cool and wet period from the middle 1940s to the middle 1970s. Overall, an increase of 1°C in mean annual temperature induced heterotrophic respiration to increase by 62 Tg C a^{-1}. In contrast to the century-scale trend from 1901 to 2000, during the last two decades (1981–2000), CO_2 fertilization was the biggest driver of the increase in soil C, while climate change alone caused soil C to decrease.

Citation: Ju, W., and J. M. Chen (2008), Simulating the effects of past changes in climate, atmospheric composition, and fire disturbance on soil carbon in Canada's forests and wetlands, *Global Biogeochem. Cycles*, *22*, GB3010, doi:10.1029/2007GB002935.

1. Introduction

[2] With the longest residence time among terrestrial carbon (C) pools [*Vogt et al.*, 1995], soil C is an important component of the terrestrial C cycle. Globally, soils contain about 1500 Pg of organic C to a depth of 1 m and a further 900 Pg from 1 to 2 m [*Kirschbaum*, 2000]. Even a small change in soil C storage could have a crucially important feedback on atmospheric CO_2 concentration and climate. Soil C dynamics is the balance between C inputs from plants and C losses from soil C decomposition, fire and erosion.

[3] Climate is one of several key factors determining the C contents of soils [*Anderson*, 1992; *Schimel et al.*, 1994]. Temperature interacts with soil water content (SWC) and other factors to affect the decomposition of soil C. The increases of soil C decomposition at higher temperatures have been overwhelmingly indicated [*Trumbore et al.*, 1996; *Davidson et al.*, 2000; *Dalias et al.*, 2001; *Sanderman et al.*, 2003; *Knorr et al.*, 2005]. Globally, soil C content and turnover time are both found to decrease exponentially with increasing mean annual temperature (MAT) [*Schimel et al.*, 1994]. This relationship is applicable to the Boreal Ecosystem-Atmosphere Study (BOREAS) area in Canada [*Peng et al.*, 1998]. However, the positive relationship between soil C content and temperature was reported by *Liski and Westman* [1997] for the 0–1 m mineral layer of forest soils in Finland and by *Callesen et al.* [2003] for well-drained Nordic forest soils (55–66°N, 6–28°E).

[4] The common perception of increased soil C emission in a warmer world was argued by some findings [*Liski et al.*, 1998; *Giardina and Ryan*, 2000; *Grace and Rayment*, 2000; *Jarvis and Linder*, 2000; *Melillo et al.*, 2002]. *Liski et al.* [1998] projected an increase in soil C storage in boreal soils under climatic warming, owing to the resistance of the decomposition of old soil organic C to climate warming. *Giardina and Ryan* [2000] analyzed published data from

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0886-6236/08/2007GB002935$12.00

82 experiments across five continents and found that decomposition rates of soil C are almost independent of MAT over the range 5–35°C by lumping soil C into a single homogeneous pool. This conclusion was late debated by *Davidson et al.* [2000] and *Knorr et al.* [2005]. They all highlighted the necessity of treating SOC as multiple pools in predicting the response of soil C to temperature increase.

[5] Global warming increases growing season length and the availability of mineral N for N-limited forests in many middle and high latitudes [*Melillo et al.*, 2002]. Therefore, it has the potential to stimulate C inputs from plants to soils to compensate at least partially for soil C loss by accelerated heterotrophic respiration. The long-term response of soil C to warming depends largely on whether the acceleration of soil C decomposition exceeds the increases of plant-derived C inputs to soils [*Davidson and Janssens*, 2006]. These positive effects should be integrated in simulating the possible response of soil C to climate change.

[6] Disagreements also exist in the response of soil C to changes in atmospheric composition. Some experiments reported detectable increases in soil C indirectly caused by elevated CO_2 [*Williams et al.*, 2000, 2004; *Jastrow et al.*, 2005]. However, some other field-scale CO_2-elevation studies concluded that the soil C increase resulting from production enhancement by CO_2 fertilization is small [*Hungate et al.*, 1997; *Tate and Ross*, 1997; *Schlesinger and Lichter*, 2001; *Gill et al.*, 2002; *Hagedorn et al.*, 2003a]. The meta-analysis of 35 independent experimental observations performed by *Jastrow et al.* [2005] reveled that CO_2 enrichment increases soil C by 5.6% even through most of these experiments have been unable to individually discern a response to CO_2 enrichment.

[7] N deposition can result in more soil C accumulation in northern forests [*De Vries et al.*, 2006]. The effects of atmospheric N deposition on organic C decomposition vary with the biochemical characteristics of plant litter. The biochemical characteristics of the dominant litter appear to modulate the effects of N deposition on soil C dynamics [*Gallo et al.*, 2005]. Observations show that N deposition retards the decomposition of soil C and hence increase soil C storage [*Hagedorn et al.*, 2003b; *Turunen et al.*, 2004; *Olsson et al.*, 2005]. *De Vries et al.* [2006] estimated an enhancement of 15 kg C kg^{-1} N on C sequestration in European forest soils by N deposition during 1960–2000.

[8] Fire strongly influences soil C storage [*Ansley et al.*, 2006]. It might significantly alter the thermal regime of the organic and mineral soil layers [*Dyrness et al.*, 1986; *Kasischke et al.*, 1995; *Harden et al.*, 2000], which, in turn, alters the decomposition rate of soil C [*Zhuang et al.*, 2002]. At a given site, during a transition to a period of more frequent disturbances, the soil C pool decreases and the soil appears to act as a source of C [*Bhatti et al.*, 2002a]. The magnitude and duration of soil C source depends on the type and severity of disturbance. The cumulative C losses after fire (prior to C accumulation) are possibly even larger than direct biomass and litter C losses from fire emissions [*Kasischke et al.*, 1995; *Zhuang et al.*, 2002]. Several studies have provided evidence of a change of soil C content with stand age [*Law et al.*, 2001; *Zhuang et al.*, 2002; *Wang et al.*, 2003; *Peltoniemi et al.*, 2004].

[9] Canada's soils contain about 262 Pg of C, approximately equivalent to 17% of the global total of 2400 Pg C to the depth of 2 m [*Kirschbaum*, 2000]. About 79% of this C content is stored in forest and wetland soils [*Siltanen et al.*, 1997; *Tarnocai*, 1998]. The largest soil C mass in Canada is stored in boreal ecozones [*Tarnocai*, 1998]. The C storage in these regions has the greatest potential to change relative to C stored in other climatic zones [*Vogt et al.*, 1995]. Many recent efforts have been made to investigate the dependence of soil C on soil texture, N availability, stand age, and drainage in Canada's forested ecosystems [*Trumbore and Harden*, 1997; *Rapalee et al.*, 1998; *Yu et al.*, 2002; *Bhatti et al.*, 2002a; *Fredeen et al.*, 2005]. A variety of models have been developed to simulate the spatial patterns of soil C and the response of soil C to climate change [*Peng et al.*, 1998; *Tarnocai*, 1999; *Liu et al.*, 2002a; *Bhatti et al.*, 2002b]. However, these models operate on landscape to regional scales and do not fully integrate the effects of climatic and atmospheric changes, vegetation, soil, disturbance, and drainage on the dynamics of soil C.

[10] The main objectives of this study are (1) to quantify historical changes in soil C contents in Canada's forests and wetlands and to analyze the spatial patterns of these changes; and (2) to partition these changes into contributions from changes in climate, atmospheric CO_2 concentration, N deposition, and fire disturbance. To achieve these objectives, six numerical simulations are conducted for the period from 1901 to 2000 using the Integrated Terrestrial Ecosystem Carbon Cycle model (InTEC) with different scenarios of input data.

2. Methods

2.1. Model Description

[11] The InTEC C model is a process-based biogeochemical model that simulates C, N, water fluxes, and pool sizes. It includes three components: (1) a modified soil C and N module adopted from CENTURY for simulating C and N dynamics [*Parton et al.*, 1987]; (2) a canopy-level photosynthesis module developed from Farquhar's leaf biogeochemical model using a temporal and spatial scaling scheme [*Farquhar et al.*, 1980; *W. J. Chen et al.*, 2000]; and (3) a three-dimensional distributed hydrological model for simulating SWC and soil temperatures [*Ju and Chen*, 2005, *Ju et al.*, 2006]. This model operates at 1 km resolution, monthly time steps (for heterotrophic respiration, net primary productivity (NPP), SWC, and soil temperature simulations) or yearly (for NPP allocation and litter transfer). In the InTEC model, major C fluxes are calculated in monthly time steps, and therefore the effects of temperature and other factors on C cycle in different seasons are well captured. In the following discussions, we use the term ''dry/wet'' and ''warm/cool'' to indicate the significant departures of the annual mean temperature and precipitation from the climatological means at a given location. The effects of their seasonal distributions on soil C sequestration are considered in the monthly modeling, but in the decadal analysis, these seasonal effects are not significant.

[12] The InTEC model progressively calculates historical annual NPP during the simulation period through a mech-

anistic aggregation of nondisturbance (climate, CO_2 concentration, N deposition, vegetation, soil texture, and drainage) and fire disturbance (stand age) effects on the C exchange between terrestrial ecosystems and the atmosphere [*W. J. Chen et al.*, 2000; *Chen et al.*, 2003], i.e.,

$$NPP_i = NPP_o \frac{F(a_i)}{F(a_o)} \prod_{j=1}^{i} \frac{2+\chi(j)}{2-\chi(j)}) \quad (1)$$

where NPP_i is the simulated NPP in year i; NPP_0 is the initial value of NPP in the starting year; $\prod_{j=1}^{i} \frac{2+\chi(j)}{2-\chi(j)})$ is the integrated effects on NPP of nondisturbance factors including climate, CO_2 fertilization and N availability [*W. J. Chen et al.*, 2000]; $F(a_i)$ and $F(a_o)$ are normalized productivities of a forest at ages a_0 (the starting year) and a_i (year i), respectively, depending on stand age and environmental conditions [*Chen et al.*, 2003].

[13] In equation (1) $\chi(j)$ is a function of climate, atmospheric CO_2 concentration, foliage N content, and SWC. Climate determines NPP through its influences on growing season length, photosynthesis rate in the growing season, soil N mineralization and fixation, and autotrophic respiration. Atmospheric CO_2 concentration is used as a surrogate of intercellular CO_2 concentration for calculating the photosynthesis rate. The ratio of intercellular to atmospheric CO_2 concentration depends on temperature, water vapor deficit, incoming solar radiation, and SWC. Total amount of N available for vegetation uptake is the summation of mineralized, deposited, and fixed N. Foliage N content is annually updated according to the amount of N uptake, N storage in biomass C pools in the previous year, and N lost with litter. The details about the calculation of $\chi(j)$ are introduced by *W. J. Chen et al.* [2000] and *Ju et al.* [2007].

[14] For each pixel, NPP_0 value in equation (1) is repeatedly tuned until the difference between NPP values from InTEC and from the daily Boreal Ecosystems Productivity Simulator model (BEPS) [*Liu et al.*, 2002b] in 1994 is smaller than a threshold, typically, ±1% between them. At the end of each year, annual NPP is allocated to the four biomass pools and some biomass C is turned over to litter C pools. NPP allocation coefficients and turnover rates of biomass C pools are prescribed for forests and wetlands separately [*Chen et al.*, 2003].

[15] Soil C and N dynamics are simulated using the modified approach of CENTURY [*Ju and Chen*, 2005; *Ju et al.*, 2006]. The major modifications include (1) the model stratifies vegetation C into four pools (foliage, stem, fine root, and coarse root) and soil C into nine pools (surface structural litter, soil structural litter, woody litter, surface metabolic litter, soil metabolic litter, surface microbial, soil microbial, slow, and passive); (2) the soil temperature effect on soil C decomposition is quantified using a modified Arrhenius-type equation where the effective activation energy for decomposition varies inversely with temperature [*Lloyd and Taylor*, 1994]; (3) the modifier for the effect of SWC on decomposition is a function of the fraction of water-filled pore space; (4) the maximum decomposition rates of slow and passive C pools are different for forests and wetlands; (5) the rate of N fixation is a function of temperature, precipitation, and the size of microbial C pool [*J. M. Chen et al.*, 2000]; and (6) temporal series of the N deposition rate for each pixel is produced on the basis of a reference map of N deposition and historical greenhouse gas emission in Canada [*Chen et al.*, 2003].

[16] The spin-up of the model is conducted using the average climatology from 1901 to 1910. Different procedures are followed to initialize C pools of forests and wetlands [*Ju et al.*, 2006]. For forests, the initialization is conducted by running the model until C dynamics reaches steady in 1900 (for stand age smaller than 100 in 2000) or in the year prior to the latest fire disturbance (for stand age larger than 100 in 2000). For wetlands, the initialization assumes the steady C dynamics for biomass and "fast" soil C pools but not for "slow soil C pools". The slow and passive soil C pools are allowed to continuously increase and their initial values are set following *Frolking et al.* [2001] with an integrated period of 100 years for the slow C pool and 8000 years for the passive pool. After the initialization of soil C pools, the model is driven by the historical climate, CO_2, and N deposition from 1901 to 2000.

[17] The change in soil C content $dC_{soil}(t)$ is updated as

$$dC_{soil}(t) = \sum_{i=1}^{4} K_{b,i} C_{b,i}(t) - \sum_{i=1}^{9} \varepsilon_i K_{s,i}(t) C_{s,i}(t) - D(t) \quad (2)$$

where $K_{b,i}$ is the turnover rate of biomass pool i; $C_{b,i}(t)$ is the size of biomass pool i in year t; ε_i is the fraction of C released to the atmosphere due to the decomposition of soil C pool i; $K_{s,i}(t)$ is the decomposition rate of soil C pool i in year t and equals the maximum decomposition rate multiplied by the scalars of soil temperature, SWC, soil texture, and litter quality (lignin content) [*Parton et al.*, 1993]; $C_{s,i}(t)$ is the size of soil C pool i; and $D(t)$ is the amount of C directly released to the atmosphere during fire disturbance.

[18] Three major types of disturbance occur in Canada's forests and wetlands (forest fire, insect-induced mortality, and timber harvest). Their relative importance changes temporally and spatially [*Kurz and Apps*, 1999]. The proportions of vegetation and soil C released to the atmosphere vary with disturbance types and severity. As spatially explicit data for these three disturbance types are not available yet, all disturbances are currently treated as fires. Fire is believed to have been the largest agent of disturbances in Canada's forests during the last century [*Kurz and Apps*, 1999]. We assume that all fire disturbances cause stand replacement, and forest regenerates in the year following a fire disturbance. The amount of C directly emitted during a fire disturbance is estimated as the sum of 100% of foliage C, 25% of woody C, and 100% of C in surface structural and metabolic pools. The remaining biomass C is transferred to corresponding soil C pools via the pathways built into the model [*Chen et al.*, 2003].

2.2. Simulation Design

[19] To achieve the goals of this study, we conduct six numerical simulations (Table 1). In Simulation I, the averages of climate data during 1901–1910, CO_2 concentration

Table 1. Description of Six Simulation Experiments

Simulation	Temperature	Precipitation	CO_2	N Deposition
I	Baseline	Baseline	Baseline	Baseline
II	Historical	Historical	Baseline	Baseline
III	Baseline	Historical	Baseline	Baseline
IV	Baseline	Baseline	Historical	Baseline
V	Baseline	Baseline	Baseline	Historical
VI	Historical	Historical	Historical	Historical

in 1900 (293 ppmv), and a N deposition rate of 0.056 g N m^{-2} a^{-1} in 1900 are used to drive the model. The changes in soil C content are caused only by fire disturbance. In Simulation II, climate changes with time while CO_2 concentration and N deposition remain at respective baseline values. This simulation is to investigate the effect of climatic change on soil C. In Simulation III, the historical precipitation and baseline values of other climate variables, CO_2 concentration, and N deposition are used to drive the model for investigating the effect of precipitation change on soil C. In Simulation IV, the atmospheric CO_2 concentration gradually increases from the baseline (293 ppmv) in 1900 to 371 ppmv in 2000, while climate and N deposition remain unchanged. This experiment is to investigate the effect of CO_2 fertilization on soil C dynamics. In experiment V, the N deposition rate of each pixel increases linearly from the baseline of 0.056 g N m^{-2} a^{-1} in 1900 to the contemporary value while CO_2 concentration and climate remain at baseline values. Its purpose is to examine the influence of N deposition on soil C. In experiment VI, we simulate the integrated effects of climate change, CO_2 concentration, and N deposition on soil C by driving the model with the historical climate, CO_2 concentration, and N deposition values. All simulation experiments use the same method to initialize C pools. Stand age changes with time after the latest fire disturbance in all simulations.

2.3. Model Validation

[20] Soil C content data complied from the Soil Landscape of Canada (SLC) data set was used to validate the ability of InTEC to capture the spatial variability of soil C. Since the SLC data set does not contain spatial information of each component inside a polygon, model outputs and SLC data were both processed to produce polygon mean values of soil C content. In total, Canada's landmass is stratified into 15,000 polygons in the SLC database. Each polygon contains several components with respective attributes, such as soil bulk density, depth, C contents of various soil layers, and area. Only 10,040 components among the total of 32,000 have soil C content values determined from measurements at sampling sites. For other components, soil C contents were estimated on the basis of the information on climate, vegetation, soil texture, and drainage class. Our model validation was first conducted for 1802 polygons with more than 50% of area containing soil C content measured at sampling sites and then for all polygons.

3. Data Used

[21] A variety of data sets were produced and compiled from various sources to drive the model (Table 2). Remotely sensed data were used for deriving leaf area index, land cover, and stand age in 2000. Forests and wetlands were differentiated on the basis of a land cover map from remote sensing in conjunction with drainage class included in SLC and wetness index derived from DEM for each polygon. There are totally 4.04×10^6 and 0.98×10^6 pixels (1 km $\times$ 1 km) of forests and wetlands. Forest stand age in 2000 was derived by combining the VEGETATION data, Canadian large-fire polygon and forest inventory databases [*Chen et al.*, 2003]. From the stand age in 2000, the time when the latest fire disturbance occurred was inferred. Gridded monthly climate data, including mean air temperature, water vapor pressure, and precipitation at 1 km resolution, were interpolated from the 0.5° global data produced by the UK Climate Research Unit from available station observations [*New et al.*, 1999, 2000]. Monthly mean radiation was estimated from monthly mean temperature range, water vapor pressure and precipitation [*Thornton and Running*, 1999]. N deposition measurements for Canada's forests

Table 2. Data Sets Used in the Simulations and Their Resources

Data Set	Source	Reference
LAI map in 1994	remote sensing	*Chen et al.* [2002]
Land cover map in 1994	remote sensing	*Cihlar et al.* [1998]
Stand age map in 2000	large forest fire scar data set, forest inventory and remote sensing	*Chen et al.* [2003]
Annual NPP in 1994	simulated by BEPS	*Liu et al.* [2002b]
Monthly mean air temperature and precipitation	interpolated from the data of UK Climate Research Unit	*New et al.* [1999, 2000]
Monthly solar radiation	simulated using air temperature range, precipitation and relative humidity	*Thornton and Running* [1999]
Annual N deposition in 1984	interpolated and extrapolated using measurements at stations	*Ro et al.* [1995]
Soil texture and soil organic C content	complied from the Soil Landscapes of Canada (SLC) database	*Shields et al.* [1991], *Schut et al.* [1994], *Lacelle* [1998], *Tarnocai* [1998]
Wetness index	derived from DEM	*Beven and Kirkby* [1979], *Beven* [1997]

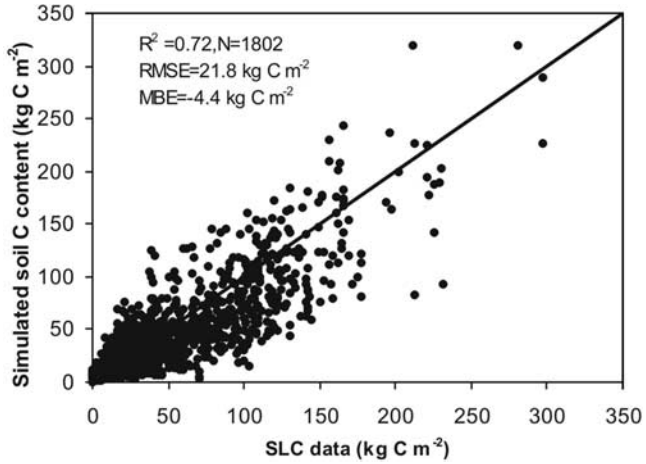

Figure 1. Comparison of simulated polygon mean soil C content with data compiled from the SLC database.

during 1983–1994 made by the Canadian Air and Precipitation Monitoring Network (CAPMN) [*Ro et al.*, 1995] were spatially interpolated and extrapolated to produce an N deposition map in 1984, from which historical annual N deposition values of each pixel were estimated. Soil texture (sand, silt and clay fractions) and drainage class were compiled from the SLC database [*Schut et al.*, 1994].

4. Results

4.1. Validation of Simulated Soil Carbon Contents

[22] The simulated polygon mean soil C contents from Simulation VI were compared with the SLC data (Figure 1). The model captures the 72% of variations of polygon mean soil C contents for 1802 polygons with more than 50% of area containing soil C contents measured at sampling sites. The root mean square error (RMSE) is 21.8 kg C m^{-2}. The model tends to slightly underestimate soil C contents, with a mean biased error (MBE) value of −4.4 kg C m^{-2} (Figure 1).

[23] The simulated results were further evaluated for all polygons. The discrepancies between simulated values and the SLC data are less than 5 kg C m^{-2} and 10 kg C m^{-2} for more than 42% and 60% of all forest and wetland areas, respectively. Large discrepancies are mainly in poorly drained areas, in which soil C contents are normally quite high and exhibit much larger variability than in well-drained areas. This model underestimation of soil C contents is mainly in the Mackenzie lowlands. The absolute difference between simulated and SLC soil C contents increases with drainage condition from excellent to very poor, implying that further efforts are required to improve the InTEC model on the initialization of C pools, parameterization, and the calculations of NPP and SWC for poorly drained areas.

4.2. Changes in Soil Carbon Contents in Canada's Forests and Wetlands

[24] In Simulation VI, total soil C increased from 160.6 Pg C to 164.5 Pg C (from 75.1 to 77.7 Pg C for forests and from 85.5 to 86.8 Pg C for forests) (Figure 2) during 1901–2000. Soils of Canada's forests and wetlands sequestrated 3.9 Pg C (2.6 Pg C in forests and 1.3 Pg C in wetlands) during the simulation period. Total soil C increased by 3.4% and 1.5% for forests and wetlands, respectively. The simulated Canada-wide average accumulation rate of soil C in 100 years was 8 g C m^{-2} a^{-1} (6 g C m^{-2} a^{-1} for forests and 14 g C m^{-2} a^{-1} for wetlands) equivalent to only 40% of the value in the northern study area of BOREAS reported by *Rapalee et al.* [1998]. The source-sink activities of soil C are determined by fire disturbance history and changes in temperature and precipitation. Soils acted as a C source of 19.6 Tg C a^{-1} during the 1910s and 1920s as a result of decreasing precipitation, increasing temperature, and reduced litter inputs from newly regenerated forests after widespread fire disturbances at the end of the 19th century and the beginning of the 20th century. Soils consistently acted as a C sink after 1943 until 1997. During these years, soil C increased by 4.6 Pg C (Figure 2). The C sink of soils shrank after 1980 because of the increase in heterotrophic respiration caused by climate warming and decreased SWC in poorly drained areas, consumption of surface litter by fires, and less litter inputs from newly regenerating forests compared with those from mature

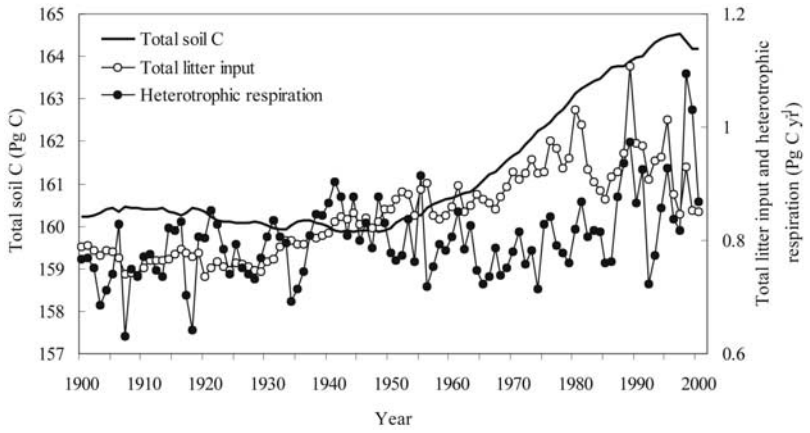

Figure 2. Simulated annual total soil carbon, total litter inputs to soils, and heterotrophic respiration in Canada's forests and wetlands in Simulation VI.

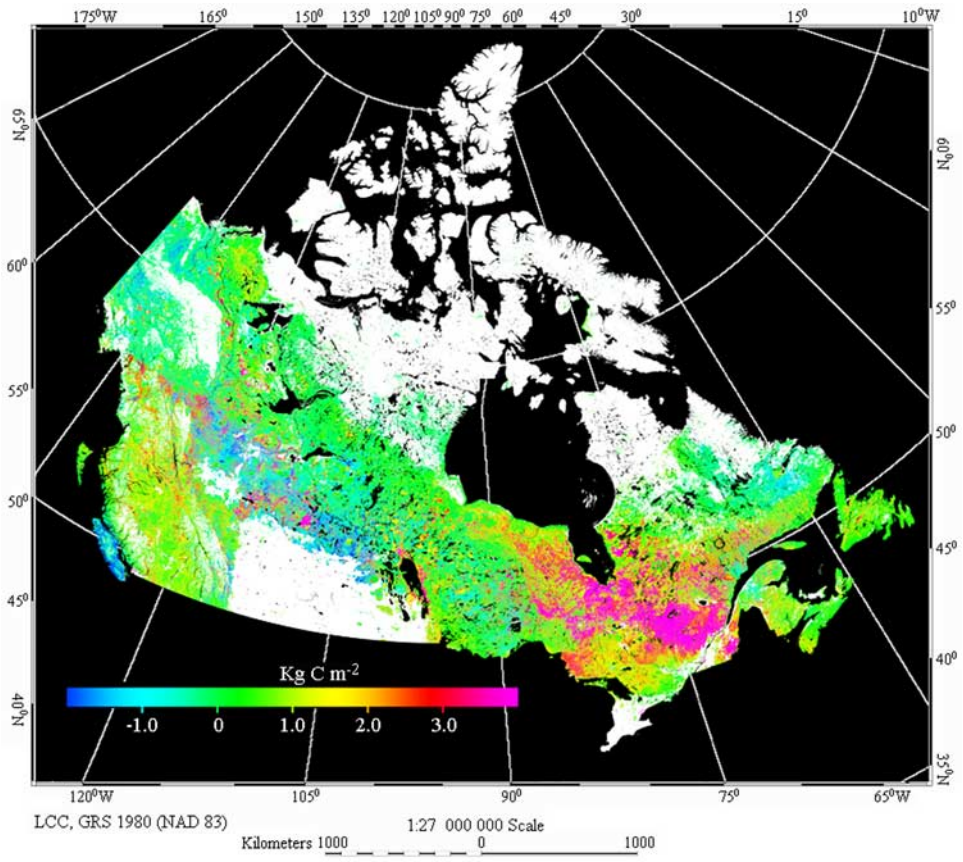

Figure 3. Spatial distribution of changes in soil C content during 1901–2000. Negative values indicate the decreases of soil C content while positive values represent the increases of soil C content during 100 simulation years.

forests (Figure 2). We assume that 100% of foliage C and 25% of aboveground woody C are directly combusted and the remaining biomass C (75% of aboveground wood, fine and coarse roots) is transferred into soil C pools during a fire disturbance process. This simple biomass-based estimation method produces the total emission intensity in close agreement with reported values by *Stocks* [1991] and *Amiro et al.* [2001]. In the year when a stand was disturbed, the content of soil C increased whereas vegetation C was totally lost. After succession, the soil continuously acted as a C source until litter inputs from vegetation were enough to compensate the heterotrophic respiratory loss.

[25] The changes of soil C were spatially heterogeneous. In the Eastern Hudson Plains, Eastern Middle Boreal Shield, Southern Boreal Shield, and Atlantic Maritime ecozones, the accumulations of soil C during 1900–2000 ranged from 2 to 4 kg C m^{-2} (Figures 3 and 4). In the forest age map (not shown here), forests in these regions were mostly at the ages ranging from 60 to 90 years in 2000. Temperature slightly decreased while precipitation increased by 15–20% during 1901–2000. The N deposition rates here were higher than those in other regions. These factors together caused the largest increase in soil C content during 1901–2000. Conversely, soil C decreased by more than 1 kg C m^{-2} in the Southern Boreal Plains ecozone. In this region, forests aged from 35 to 55 years in 2000 and were in the early stage of accumulating soil C. Temperature increased above 2°C while precipitation decreased by 5–10% during 1901–2000. This combination of fire disturbance history and changes in temperature and precipitation induced the highest loss of soil C.

[26] Figure 5 shows the temporal patterns of soil C contents in various ecozones. Total C contents increased in most ecozones. The exception is in the Taiga Cordillera ecozone, where soil C content decreased by 1.5% (0.03 Pg C). The soil C contents increased by about 10% in Southern Boreal Shield (0.8 Pg C) and more than 5% in the Montane Cordillera, Middle Boreal Shield, and Mixedwood Plains ecozones. In other ecozones, the increases of soil C were below 4%. The temporal patterns of soil C varied distinctly in different ecozones. The soil C contents decreased prior to the middle 1950s in Taiga Cordillera and prior to the middle 1940s in the Boreal Plains, Boreal Cordillera, and Western Boreal Shield ecozones. The decreases in soil C contents in the 1980s and 1990s were pronounced in the Southern Arctic, Taiga plains, Taiga Cordillera, Taiga Shield, Boreal Cordillera, and Western Boreal Shield ecozones, and less pronounced in the Montane Cordillera, Hudson Plains, Middle Boreal Shield, Southern Boreal Shield, and Eastern Boreal Shield ecozones.

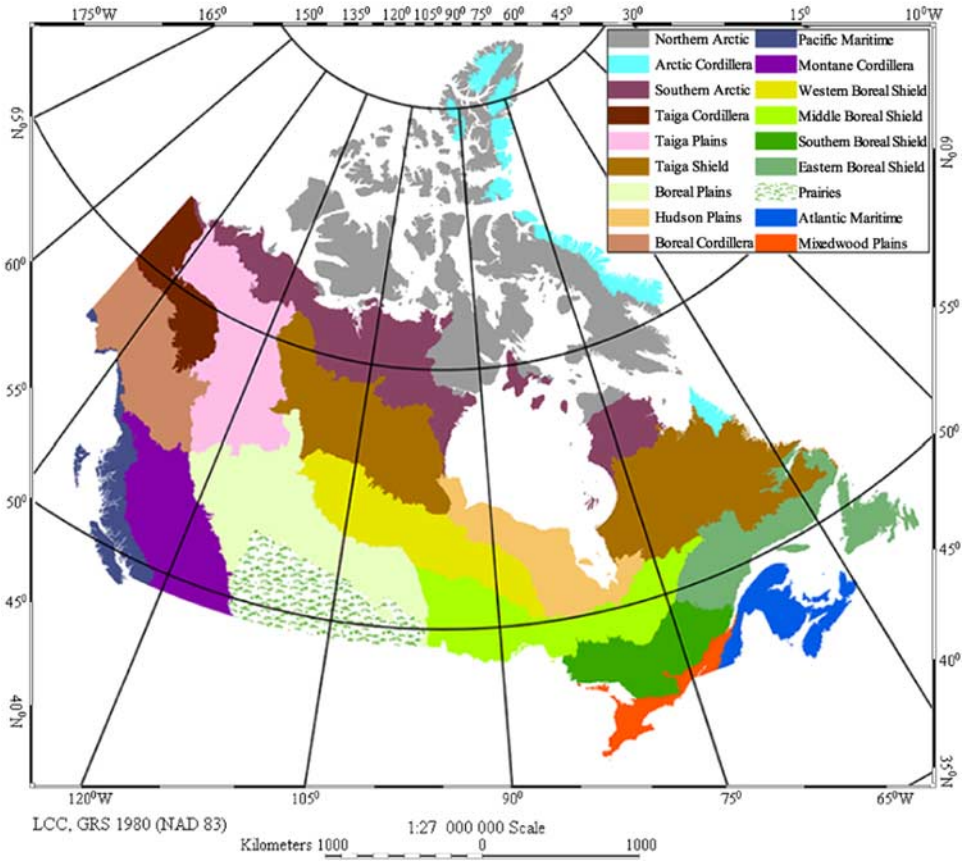

Figure 4. Spatial distribution of various ecozones in Canada.

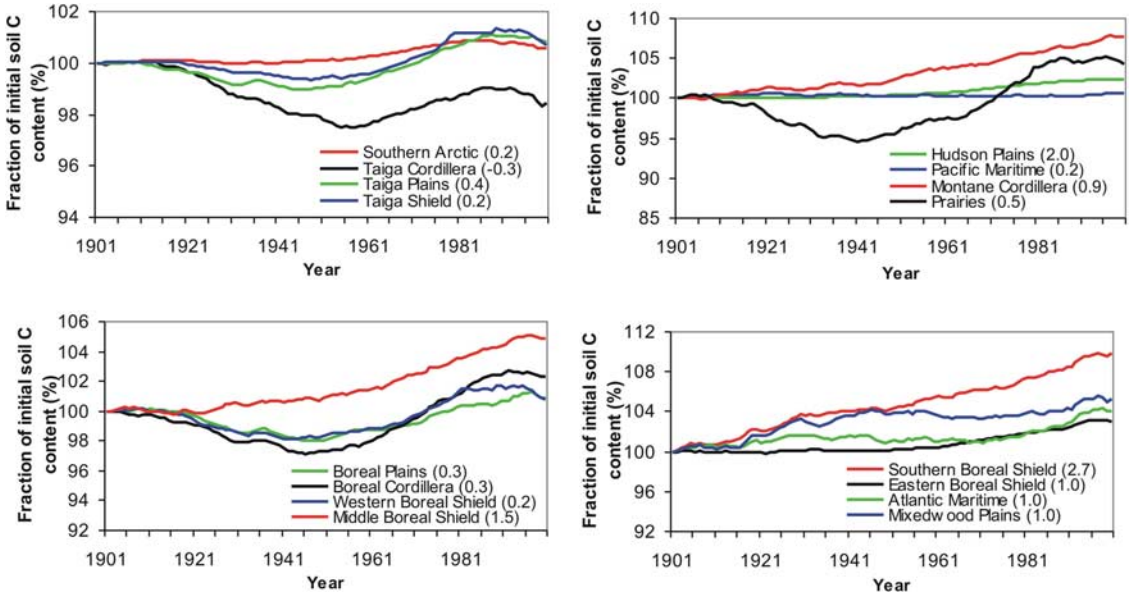

Figure 5. The relative temporal changes in soil C content in various ecozones. The values shown are the ratios of soil C contents at a given time to the initial values in 1900. Values over 100 mean that soil C contents increase relative to the initial values in 1900.

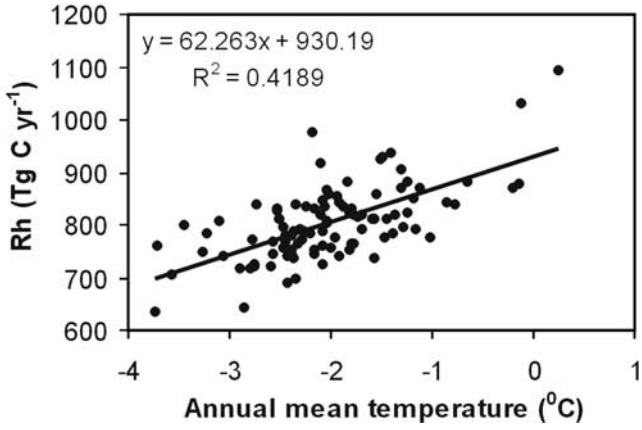

Figure 6. The relationship between annual total heterotrophic respiration and annual mean air temperature in Canada's forests and wetlands. An increase of 1°C in annual mean air temperature causes total heterotrophic respiration to increase by 62 Tg C a^{-1}.

4.3. Heterotrophic Respiration in Canada's Forests and Wetlands During 1901–2000

[27] In Simulation VI, simulated annual total soil heterotrophic respiration increased gradually from an average of 740 Tg C a^{-1} in the 1910s to 890 Tg C a^{-1} during 1991–2000 owing to the climate-induced increases in the soil C decomposition rates and increasing soil C contents (Figure 2). MAT is the key determinant of annual heterotrophic respiration and explains 42% of interannual variability in total heterotrophic respiration (Figure 6). On average, an increase of 1°C in MAT enhanced total heterotrophic respiration by 62 Tg C a^{-1}. At the continental scale, the simulated annual heterotrophic respiration was found to be linearly correlated with MAT, but not with precipitation, because of the different responses of heterotrophic respiration to precipitation variations in upland forests and poorly drained forests and wetlands. The increase of precipitation might retard heterotrophic respiration in poorly drained wetlands and forests and stimulate heterotrophic respiration in upland forests. Our dependences of simulated annual heterotrophic respiration on temperature and precipitation at the continental scale are similar to the modeling results of *Tian et al.* [1999] using the TEM model [*Raich et al.*, 1991] for the conterminous U.S. and to the findings of *Raich et al.* [2002]. The compilations of in situ soil respiration measurements by *Raich et al.* [2002] showed that monthly soil CO_2 emission is exponentially related to temperature at individual sites with different climate conditions. The relationship between estimated annual global soil respiration and MAT over land is linear. There is no correlation between estimated annual global soil CO_2 emissions and global precipitation [*Raich et al.*, 2002].

[28] Total heterotrophic respiration exhibited notable interannual and decadal variations (Figure 2). In Simulation VI, the mean annual heterotrophic respiration increased gradually from 770 Tg C a^{-1} in 1901 to an average of 870 Tg C a^{-1} during 1938–1942 and then decreased until the late 1960s in response to the slight decrease in temperature and increases in soil wetness. In the last two decades, soil heterotrophic respiration increased rapidly owing to the widespread temperature increases as well as the decreases in SWC in poorly drained forests and wetlands. However, in years 1992 and 1993 after Mt. Pinatubo eruption, heterotrophic respiration rates were 724 Tg C a^{-1} and 774 Tg C a^{-1}, respectively. They were 161 Tg C a^{-1} and 111 Tg C a^{-1} lower than the decadal average. This is consistent with the findings of *Jones and Cox* [2001]. The highest heterotrophic respiration was 1100 Tg C a^{-1} in 1998. From 1901 to 2001, the standard deviation (SD) of annual heterotrophic respiration was 71 Tg C a^{-1}, smaller than the SD of NPP which was 97 Tg C a^{-1}. Although both heterotrophic respiration and NPP immediately respond to climate and other variables, the NPP variation also induces the variation in the transfer of organic matter to the soil, causing delayed and dampened heterotrophic variation. The interannual variability of heterotrophic respiration is primarily caused by climate. In Simulation II, the SD value of annual heterotrophic respiration was 63 Tg C a^{-1} while the SD value of annual NPP was 80 Tg C a^{-1}. Fire disturbance caused notable decadal trend, but less interannual variability in heterotrophic respiration. In Simulation I driven by the baseline values of all nondisturbance factors, the SD value of annual heterotrophic respiration was only 2 Tg C a^{-1} (the SD value of NPP was 5 Tg C a^{-1}). Total annual heterotrophic respiration declined from the beginning to a minimum of 760 Tg C a^{-1} in 1916 because of frequent fire disturbances at the end of the 19th century and the first decade of the 20th century. With the accumulation of soil C, then annual heterotrophic respiration gradually increased to a peak value of 790 Tg C a^{-1} in 1981 and slightly decreased after 1991 with the decrease in soil C caused by more fire disturbances.

4.4. Effect of Fire Disturbance on Soil Carbon Content

[29] In Simulation I, only stand age changes progressively after the last fire disturbance while other forcing factors remain at the baseline values. The change in soil C is due to only the effect of fire disturbance. In the first two and the last three decades of the last century, fire disturbance rates were above normal. From 1930 to 1970, fire disturbance occurred at a low frequency. For simplicity, we used a fire disturbance factor to include both its negative (direct carbon emission during fire disturbance) and positive (regrowth after fire disturbance) effects on C balance. The total soil C content slightly decreased from 1901 to the early 1940s and then gradually increased until the early 1990s. It declined again in response to higher fire disturbance rates. The increase in total soil C content in Canada's forests and wetlands resulting from stand age dynamics caused by fire disturbance was 1 Pg C during 1901–2000.

4.5. Effects of Nondisturbance Factors on Soil Carbon

[30] The total increase of soil C induced by nondisturbance factors in Canada's forests and wetlands was 2.93 Pg C (2.17 Pg C for forests and 0.76 Pg C for wetlands) (Table 3). The average enhancement on soil C by nondisturbance factors was 6 g C m^{-2} a^{-1}over 100 simulation years (5 g C m^{-2} a^{-1} for forests and 8 g C m^{-2} a^{-1} for wetlands).

Table 3. Effects of Nondisturbance Factors on Total Soil C Content of Canada's Forests and Wetlands in the Last Century (1901–2000) and During 1981–2000[a]

	1901–2000 ΔC_{soil} Pg C			1981–2000 ΔC_{soil} Pg C		
	Forests	Wetlands	Total	Forests	Wetlands	Total
Climate change	0.83	0.46	1.29	−0.27	0	−0.27
Precipitation change	0.87	0.25	1.12	−0.03	0.06	0.03
CO_2 fertilization	0.60	0.16	0.76	0.25	0.06	0.31
N deposition	0.63	0.09	0.72	0.26	0.04	0.30
All nondisturbance factors	2.17	0.76	2.93	0.28	0.13	0.41

[a]Nondisturbance factors are N deposition, CO_2 fertilization, and climate change (climate influences growing season climatic conditions, growing season length, net N mineralization, and N fixation). The sum of the individual effects differs from the integrated total because of the interactions among these components. ΔC_{soil} is departures of soil C content simulated with changing climate, CO_2, and N deposition from those simulated with constant nondisturbance factors.

The largest contribution was from precipitation change (1.12 Pg C), followed by CO_2 fertilization, and N deposition. The average accumulation rates of soil C related to climate change were 2 g C m^{-2} a^{-1} for forests and 5 g C m^{-2} a^{-1} for wetlands, respectively. Similar to *Potter et al.* [1993] and *Cao and Woodward* [1998], this model includes the control of N availability over the decomposition of soil C. N deposition as an external input to soils reduced the frequency of low N availability limiting soil C decomposition and increased the fraction of litter C transferred to metabolic pools with higher decomposition rates. This factor made the smallest contribution to the increase in total soil C content although it made the second largest contribution to the increases in NPP and biomass C from 1901 to 2000. The total input of annual N deposition averaged over the period during 1901–2000 was 0.74 Tg N, increasing from 0.28 Tg N in 1901 to 1.19 Tg N in 2000. The enhancement of N deposition on soil C sequestration was simulated to be 0.72 Pg C during 1901–2000 (Table 3). Our simulated impact of N deposition on the net soil C sequestration is 17 kg C kg^{-1} N during 1901–2000, 13% higher than the value of 15 kg C kg^{-1} N estimated by *De Vries et al.* [2006] for European forests, but 19% lower than the estimate of 21 kg C kg^{-1} N by *Nadelhoffer et al.* [1999] for temperate forests.

[31] CO_2 fertilization resulted in NPP increase. In addition, it increased C inputs to soil C pools by an extent enough to increase C:N ratios and to limit decomposition. Therefore, CO_2 fertilization resulted in a greater increase in soil C accumulation compared with N deposition. The effect of CO_2 fertilization increased soil C sequestration by 0.5% (0.76 Pg C) relative to the initial value during 100 years. The average enhancement on soil C by CO_2 fertilization was 2 g C m^{-2} a^{-1} over 100 years, much smaller than the value of 40 g C m^{-2} a^{-1} for 5–8 years in an elevated CO_2 experiment reported by *Jastrow et al.* [2005]. This implies that the effect of CO_2 fertilization on soil C sequestration shrank with time because of N limitation.

[32] Compared with CO_2 and N deposition, climatic change gave rise to relatively large interannual and decadal variations in total soil C content (Figure 7). The enhancement of climatic change on soil C accumulation occurred mainly in the cool and wet period from 1943 to 1977. Climate change had a very small positive and even negative effect on soil C accumulation in periods from 1920 to 1931, from 1937 to 1943, and from 1981 to 1998, when climate was warm and dry. Climate change caused soil C to decrease, although it was still an important contributor to the increase in NPP during 1981–2000. These results suggest that more extensive field and modeling efforts are needed to project the fate of soil C under the likely future climate change.

[33] The effect of CO_2 fertilization on soil C has very small spatial variability since a single CO_2 value is used in each year for all pixels. The increase in soil C caused by N deposition exhibits a clear decreasing gradient from industrialized to rural areas in association with the spatial distribution pattern of N deposition. However, the climate-induced changes in soil C contents vary largely in different areas. The climate-induced increases in soil C are in eastern Canada, mainly in the Hudson Plains, Middle, Eastern, and Southern Boreal Shield, Mixedwood Plains and Atlantic Maritime ecozones, ranging from 2% to 6%. Climate induced the highest increase in soil C content in the Southern Boreal Shield ecozone. Climate resulted in decreases in soil C in western Canada, especially in the Pacific Maritime (2%) and Taiga Cordillera (1%) ecozones. Climate caused successive decrease in soil C from the middle 1900s to the middle 1940s in the Taiga Cordillera, Taiga Plains, Boreal Plains, and Boreal Cordillera ecozones.

5. Discussion

[34] Because of the integrated effects of fire disturbance and nondisturbance factors, soils of Canada's forests and wetlands were overall a C sink during 1901–2000. The simulated average accumulation rate of soil C was 8 g C m^{-2} a^{-1} (6 g C m^{-2} a^{-1} for forests and 14 g C m^{-2} a^{-1}

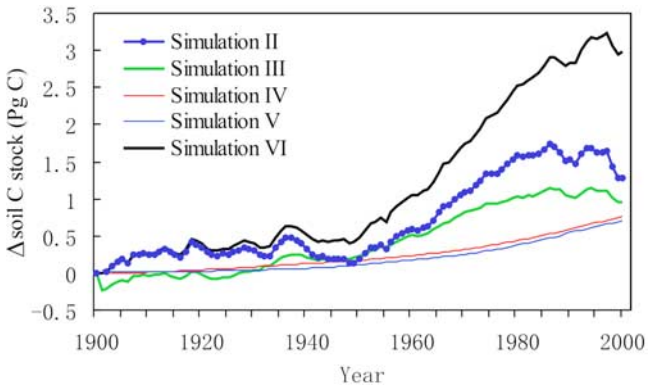

Figure 7. Partition of the accumulated effects of nondisturbance factors on total soil C content of Canada's forests and wetlands from 1901 to 2000. The values presented here are the differences between the total soil C contents output from Simulations II to IV and the total soil C content simulated in Simulation I. Positive values represent that nondisturbance factors have enhancement effects on soil C content. The definition of each Simulation is given in Table 2.

for wetlands). *Gorham* [1995] and *Rapalee et al.* [1998] estimated that northern wetlands of Canada had been persistent C sinks, ranging from 20 to 30 g C m^{-2} a^{-1}over over the 5,000 to 100,000 years. Caution should be taken for the compassion between our and their estimates. Our value is only for the increase of soil C while their estimates include C sequestration by vegetation and soil. Of course, the C sequestration in soil accounts for the most of the long-term total C sinks.

[35] In the 1980s and 1990s, North America was estimated as a carbon sink by inventory data, bottom-up modeling, and atmospheric inversion [*Fan et al.*, 1998; *Houghton et al.*, 1999; *Chen et al.*, 2003; *Potter et al.*, 2003; *Denman et al.*, 2007]. However, the disagreement about the magnitude of this C sink still exists. This study shows that Canada's forests and wetlands were a C sink of 0.07 Pg C a^{-1} (with 0.06 Pg C a^{-1} going to soils and 0.01 Pg C a^{-1} stored in vegetation) for the period from 1982 to 1999, accounting for 23% to 35% of the C sink (0.2 to 0.3 Pg C a^{-1}) in North America reported by *Potter et al.* [2003] on the basis of the CASA model and remote sensing data. The C sink in Canada's forests and wetlands during 1988 to 1992 estimated in this study was 0.05 Pg C a^{-1}, equals to 14% to 33% of the C sink (0.15–0.3 Pg C a^{-1}) in the United State estimated by *Houghton et al.* [1999].

[36] The integrated effects of nondisturbance factors resulted in a larger increase in the total soil C content than did fire disturbance factor during 1901 to 2000. Nondisturbance factors together caused the total soil C to increase by 2%. The magnitude of simulated soil C increase in Canada's forests and wetlands is close to the model results by *Jones et al.* [2005] They estimated that total soil C increased by 2% for regions in 50–90°N during 1860 to 2000.

[37] Climate was found to be the largest contributor to the total soil C accumulation during 1901–2000. The total increase in soil C content due to climate change was 1.29 Pg C (0.83 Pg C for forests and 0.46 Pg C for wetlands). Estimated response of soil C to climate change depends largely on the temperature sensitivity of soil C decomposition (Q_{10}) used in the model. Laboratory incubation and in situ field studies have produced a wide range of Q_{10} values [*Lenton and Huntingford*, 2003]. In this study, the reaction of the decomposition of all soil C pools to temperature is described using the modified Arrhenius-type equation where the effective activation energy for respiration varies inversely with temperature, namely Q_{10} decreases with temperature. However, this equation is still empirical and may not fully describe acclimation of soil organic C decomposition to long-term temperature increase. The mechanism of this acclimation is still not clear in reported studies. Through a 1-year soil warming experiment, *Luo et al.* [2001] showed that both the temperature sensitivity and the rate of decomposition decreased under warming conditions, but it is unclear if these decreases are due to substrate labile carbon limitation or the impact of warming on microbes. *Giardina and Ryan* [2000] suggested that decomposition rates of organic C in mineral soil do not vary with temperature. If this suggestion is true, Arrhenius-type equation might overestimate soil C release and our estimate of the contribution of climate change to soil C accumulation in the past 100 years might be too low. However, the conclusion of *Giardina and Ryan* [2000] was argued by *Davidson et al.* [2000]. They criticized the methodological factors and assumption that soil C exists as a single homogenous pool used by *Giardina and Ryan* [2000]. Therefore, more efforts are needed to use the process-based methodology instead of the empirical ones to reliably represent the temperature sensitivity of soil C decomposition in the model for estimating the feedback of the terrestrial C cycle to global warming.

[38] Canada's forests and wetlands store 164.5 Pg C in 2000 based on our estimate in this study, accounting for about 7% of the global total of 2400 Pg C to the depth of 2 m [*Kirschbaum*, 2000]. Heterotrophic respiration was estimated as 0.9 Pg C a^{-1} during 1991 to 2000. This value is a very small fraction of global soil respiration during 1980–1994 estimated by *Raich et al.* [2002]. This is mainly due to low soil C decomposition rates under conditions of low temperature and water logging in poorly drained areas. An increase of 1°C in MAT was estimated to increase heterotrophic respiration by 0.06 Pg C a^{-1}. In the future, with global warming and corresponding reduction in SWC, more soil C will be possibly respired from the soils of Canada's forests and wetlands to the atmosphere. The simulated significant increase of heterotrophic respiration in the warm and dry year 1998 implies this possibility.

[39] Because of the lack of spatial data to differentiate disturbance types, all disturbances were treated as fire disturbance in this study. The direct C release during a fire was estimated using a simple biomass-based model and the effect of fire severity has not been included. These simplifications could induce errors in the estimates since changes in disturbance type and fire severity affect direct C loss and subsequent decomposition and forest regeneration. In addition, regeneration was assumed to occur in the year after disturbance. This may not be the case in some locations. The uncertainties in the estimate of soil C dynamics related to disturbance may be constrained with the development of new remote sensing algorithms and sensors.

[40] Heterotrophic respiration is sensitive to SWC as indicated by *Denman et al.* [2007]. This model is driven by the climate data set produced from station observations. In Canada's northern areas, historical climate measurements are relatively sparse. The uncertainties in climate data, especially precipitation, might result in some uncertainties in simulated soil C changes. N deposition is an important source of N available for forests and wetlands in high latitudes where N mineralization is normally low. This study shows that N deposition made a contribution to soil C increase in Canada's forests and wetlands compatible to CO_2 fertlization. However, observations of N deposition are still sparse. Increasing the availability of N deposition data will definitely increase our confidence on the estimate of the terrestrial C cycle.

6. Conclusions

[41] This paper reports the responses of soil C in Canada's forests and wetlands to changes in climate, atmospheric

composition, and fire disturbance through a set of model simulations designed to separate these effects. The major findings are

[42] 1. The total of simulated soil C in Canada's forests and wetlands increased from 160.6 Pg C to 164.5 Pg C during 1901–2000, with total increases of 2.6 Pg C (3.4%) and 1.3 Pg C (1.5%) in forests and wetlands, respectively. During the 100 year simulation period, the total accumulations of soil C ranged from -2 kg C m^{-2} to 4 kg C m^{-2}, varying spatially depending on forest age structure and the temporal patterns of changes in temperature and precipitation. Precipitation was found to be the biggest contributor to the total 2.93 Pg C increase in soil C caused by nondisturbance factors from 1901 to 2000 (2.17 Pg C for forests and 0.76 Pg C for wetlands), followed by CO_2 fertilization, and N deposition.

[43] 2. Climate caused sizable interannual variability in NPP and heterotrophic respiration and notable decadal change in total soil C content. In Simulation II, the SD value of NPP was 82 Tg C a^{-1}, larger than that of 63 Tg C a^{-1} of heterotrophic respiration, implying that NPP was more sensitive to climate change than heterotrophic respiration. Climate-induced enhancement of soil C accumulation occurred primarily during the cool and wet period from 1943 to 1977. During dry and warm periods, climate had a very small positive, and even negative, influence on soil C accumulation. If only the effects of climate change were considered, soils of Canada's forests and wetlands would have lost 0.27 Pg C during the dry and warm period from 1981 to 2000.

[44] 3. Disturbance directly releases C to the atmosphere and changes productivity, consequently affecting soil C storage and dynamics. In the 100 simulation years, the total soil C content in Canada's forests and wetlands increased by 1 Pg C because of the net effect of stand disturbance processes.

[45] It is still impossible to completely validate our estimates for Canada's forests and wetlands. However, partial validation by comparing simulated soil C content in year 2000 with the SLC data set shows that the model demonstrates the ability to capture the spatial variability of soil C. The model simulated an increase of 2% in total soil C from 1901 to 2000 induced by nondisturbance factors, and this result is close to the model estimate of *Jones et al.* [2005] for regions in 50–90°N during 1860 to 2000. The average soil C accumulation rate in wetlands was estimated to be 14 g C m^{-2} a^{-1} during 1901 to 2000, smaller than the estimated C sequestration of 20 to 30 g C m^{-2} a^{-1} by Canada's northern wetlands over the past 5,000 to 100,000 years [*Gorham*, 1995; *Rapalee et al.*, 1998]. Our estimates shows that Canada's forests and wetlands were C sink in the 1980s and 1900s, compatible to the estimates for the United States by *Houghton et al.* [1999] using forest inventory and land use change data and for the North America by *Potter et al.* [2003] using satellite and the CASA model. The change of soil C is the balance between litter inputs and heterotrophic respiration. The responses of soil C to temperature and soil water content (SWC) are complex [*Denman et al.*, 2007]. Most soils of Canada's forests and wetlands are normally wet. The reduction of SWC accompanying temperature elevation significantly enhances heterotrophic respiration and causes soil C to reduce. The negative effect of climate change on soil C was suppressed by the positive effect of CO_2 fertilization and N deposition during 1981–2000. The average 10-year enhancement of soil C by the integrated effect of climate, CO_2 fertilization and N deposition was 0.2 Pg C during 1981 to 2000 while this value was 0.4 Pg C during 1901 to 1980. In the future, with the saturation of CO_2 fertilization, continuing increase of temperature, and decrease of SWC, soils of Canada's forests and wetlands might sequester less or even loose C. This is similar to the global results from the C^4MIP project [*Denman et al.*, 2007].

[46] As discussed above, there are uncertainties in simulated soil C dynamics, mainly related to the quantification of temperature sensitivity of soil C decomposition, estimation of direct C release by different types of disturbances, and reliability of historical climate and N deposition data in northern areas. Further efforts are needed to address these issues.

[47] **Acknowledgments.** This research is part of the Fluxnet-Canada Research Network (FCRN) sponsored by the Canadian Foundation for Climate and Atmospheric Sciences, Natural Science and Engineering Council of Canada, and BIOCAP Canada. We are indebted to Josef Cihlar of the Canada Centre for Remote Sensing for the use of the Canada-wide land cover map, Agriculture and Food Canada for the Soil Landscape of Canada. We also gratefully acknowledge constructive comments from two anonymous reviewers and editors, which helped improve the quality of the manuscript greatly.

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