High sensitivity of peat decomposition to climate change through water-table feedback

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Historically, northern peatlands have functioned as a carbon sink, sequestering large amounts of soil organic carbon, mainly due to low decomposition in cold, largely waterlogged soils[1,2]. The water table, an essential determinant of soil-organic-carbon dynamics[3–10], interacts with soil organic carbon. Because of the high water-holding capacity of peat and its low hydraulic conductivity, accumulation of soil organic carbon raises the water table, which lowers decomposition rates of soil organic carbon in a positive feedback loop. This two-way interaction between hydrology and biogeochemistry has been noted[3,5–8], but is not reproduced in process-based simulations[9]. Here we present simulations with a coupled physical–biogeochemical soil model with peat depths that are continuously updated from the dynamic balance of soil organic carbon. Our model reproduces dynamics of shallow and deep peatlands in northern Manitoba, Canada, on both short and longer timescales. We find that the feedback between the water table and peat depth increases the sensitivity of peat decomposition to temperature, and intensifies the loss of soil organic carbon in a changing climate. In our long-term simulation, an experimental warming of 4 °C causes a 40% loss of soil organic carbon from the shallow peat and 86% from the deep peat. We conclude that peatlands will quickly respond to the expected warming in this century by losing labile soil organic carbon during dry periods.

Accumulation of organic soil in a peatland increases its soil water-holding capacity and raises the water table[3,5–8]. The higher water table lowers soil organic carbon (SOC) decomposition rates by creating anoxic conditions, and the peatland further grows in depth. Insulation by peat tends to keep summertime soil temperatures relatively low, further decreasing SOC decomposition rates[11,12]. Scientists have long known of this positive feedback process, often called paludification (pond-making)[3,5–8,13]. Paludification has been considered when explaining past or current peatland formation under favourable environmental conditions for such processes[5,13]. However, any positive feedback process is inherently a destabilizing forcing that increases rates of change[14], and the water–peat feedback should intensify peat decomposition when environmental conditions become unfavourable.

While a few studies have considered paludification using analytical models that include feedback between soil depth change and hydrology[3,6,7], fast-timescale processes, especially energy dynamics, have not been incorporated due to limitations of analytical models. Development of an integrated process-based simulation model of peatland biogeochemistry has been difficult because of the challenges of simulating physical aspects such as soil hydrology and thermal dynamics and the lack of computational and conceptual modelling frameworks[9]. In this study, we couple the land-surface model ED-RAMS[15,16] (Ecosystem Demography Model version 2 integrated with the Regional Atmospheric Modeling System[17]) with a peatland biogeochemical model to closely reproduce fine-timescale dynamics of peatland physical conditions and simulate long-term effects of the soil physics–peat feedback. This model can reproduce theoretical responses of analytical models from vertically explicit physical treatment of soil hydrological and energy dynamics. The model can also show the transient responses of peatland to climate change. SOC is compartmentalized into metabolic, structural and humic pools with various turnover times (see Supplementary Information, Methods). Litter inputs are partitioned to metabolic and structural SOC pools according to lignin fraction[2,18], and partially decomposed structural SOC is transferred into the humic pool. Metabolic and structural pools comprise the young, less dense peat (fibrous layer), and the humic pool comprises the old, dense peat (humic layer). This multipool structure reproduces the dynamic SOC quality distribution and SOC–hydrology feedback[3–10,19].

Our model captures dynamics of the water table and soil temperature in 2003 at the Old Black Spruce (OBS) site of the Northern Study Area of the Boreal Ecosystem–Atmosphere Study (BOREAS)[4,20,21] (55.9° N and 98.5° W, 259 m above sea level). This site has shallow peat layers (Fig. 1) and is mostly ombrotrophic and disconnected from regional hydrology. Simulated water-table depth and soil temperature generally compare favourably with field observations. The spring ice melt is in early May, and the gradual drawdown of the water table ends in mid-June. During summer, the water table is in a dynamic steady state, with sudden wetting due to major rainfall events and gradual drying between rainfalls. Seasonal soil temperatures are somewhat underestimated during early summer and overestimated in late summer, but the general trend is correctly reproduced. Disturbances in soil temperature due to summer precipitation are successfully captured. This

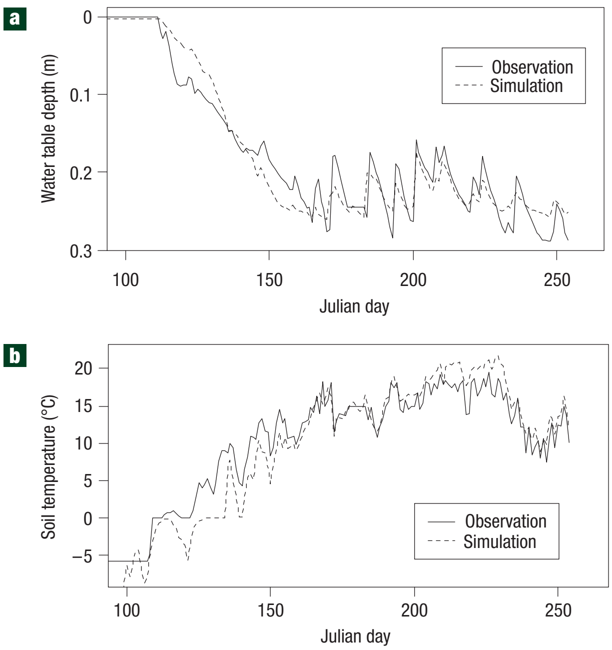

Figure 1 Comparisons of the simulation and the field observation in year 2003 at BOREAS OBS. **a**, Water-table depth. **b**, Temperature of the fibrous peat layer. The observed data were recorded at a poorly drained, hollow subsite.

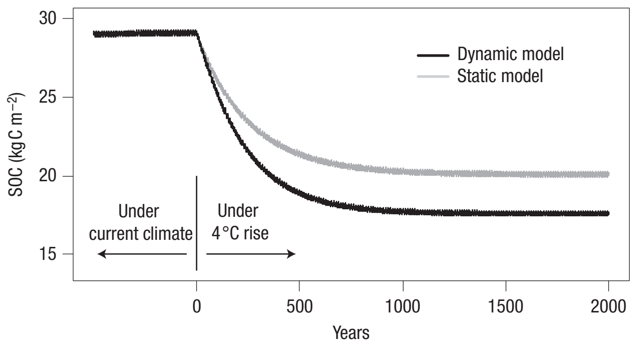

Figure 2 Changes in SOC equilibria due to an experimental warming of 4 °C at BOREAS OBS. The equilibrium SOC under the warmer climate is lower in the dynamic soil-depth model (dynamic model) than in the static soil-depth model (static model) because positive feedbacks from reduced water retention and summertime insulation increase the temperature sensitivity of peatland decomposition.

reasonable reproduction of soil physical properties suggests that our model can capture the hydrological and thermal dynamics of peatlands (more examinations of model performance are provided in Supplementary Information, Methods).

Since SOC decomposition in the model is controlled by soil moisture and temperature, the ability of our model to simulate the soil physical characteristics allows us to realistically calculate decomposition rates for a given time and vertical positions in soil using temperature- and moisture-dependence functions (see Supplementary Information, Fig. S1). To explicitly capture the long-term feedback between soil physical conditions and SOC dynamics, the peat depths in our simulation model are continuously updated using the amount of SOC, peat density and carbon fraction. Variables to calculate soil physics (for example total soil water-holding capacity and diffusion rates of moisture and heat) are also updated accordingly, thereby reproducing the realistic soil conditions that control decomposition rates.

The equilibrium SOC from ED-RAMS (29.1 kg C m^{-2}, 0.33 m in depth) closely matches the best estimate of a multi-isotope model[22] (28.9 kg C m^{-2}, 0.35 m). We then force the model with an instantaneous temperature rise of 4 °C to observe the temperature sensitivity of the model (Fig. 2). To illustrate the effect of the soil physics–carbon feedback process of the dynamic model, the result is compared against that of the static model in which soil depths are constant throughout the simulation. Before year 0, the models are in equilibria under the current climate regime. However, with warming of 4 °C at simulation year 0, the dynamic and static models lose SOC and eventually establish new equilibria under the warmer climate. The dynamic model shows stronger sensitivity to warming, losing 40% of SOC from the current equilibrium ($Q_{10} = 3.6$). The static model, which loses 31% SOC, has a Q_{10} value of 2.5, which is within the range shown by conventional biogeochemical models[18,23]. Note that the intrinsic temperature sensitivity of decomposition in our model is $Q_{10} = 2.0$ (see Supplementary Information, Fig. S1). In the static model, the drier conditions accompanying the warming (for example increase in evapotranspiration) raise Q_{10} from 2.0 to 2.5. In the dynamic model, the warming accelerates SOC decomposition, and soil depth decreases. This loss in water-storage capacity further dries the peat column.

Next, we apply the model to the BOREAS Fen site[20,24], which is 7 km northeast of the OBS site and has much deeper peat. The Fen site is located in a small catchment and is more affected by regional hydrology. Unlike the OBS site, the Fen site has large seasonal fluctuations in its water table[5]. Due to its geomorphology, the Fen site is often ponded for an extended duration, especially in spring and early summer. To capture these hydrological characteristics, we turned off the fast-surface-runoff process (see Supplementary Information, Methods) to allow for extended ponding in simulations of the Fen site (see Supplementary Information, Fig. S3). Despite the lack of a lateral water flow process in our one-dimensional water-balance model, the model sufficiently captures the hydrological dynamics of the Fen site.

Because of the higher water table, the Fen site accumulates large amounts of peat (108 kg C m^{-2}, 2.1 m in depth) under the current climatic regime, after a simulation of 2000 years (Fig. 3). We then forced the Fen equilibrium with an instantaneous temperature rise of 4 °C at year 2000 and observed the transient and equilibrium responses to the warming. In years 2000–2200, the decomposition of the structural SOC increases the input to the humic SOC, and the latter pool transiently increases in depth. Thus, the initial peatland response to the warming is relatively slow, showing some resistance to peat loss. In this phase, SOC loss from the humic layer due to its temperature sensitivity is smaller than the increase in the input from the fibrous layer. A few centuries after the initiation of warming, the SOC amount in the fibrous layer becomes stabilized under the new climate regime, and the input to the humic pool becomes low. This phase (2300–2600) has the maximum SOC loss rate, and the water table is periodically drawn down to the humic layer (Fig. 3b), further intensifying the decomposition of the deeper peat. Eventually the system reaches a new equilibrium of approximately 17 kg C m^{-2} under the warmer climate regime. The loss of 86% of SOC from the previous equilibrium due to the 4 °C warming represents a Q_{10} of 136. This certainly is not enforced by the temperature dependence of SOC decomposition itself ($Q_{10} = 2.0$); rather, most of the SOC loss is caused by the positive feedback between water loss and

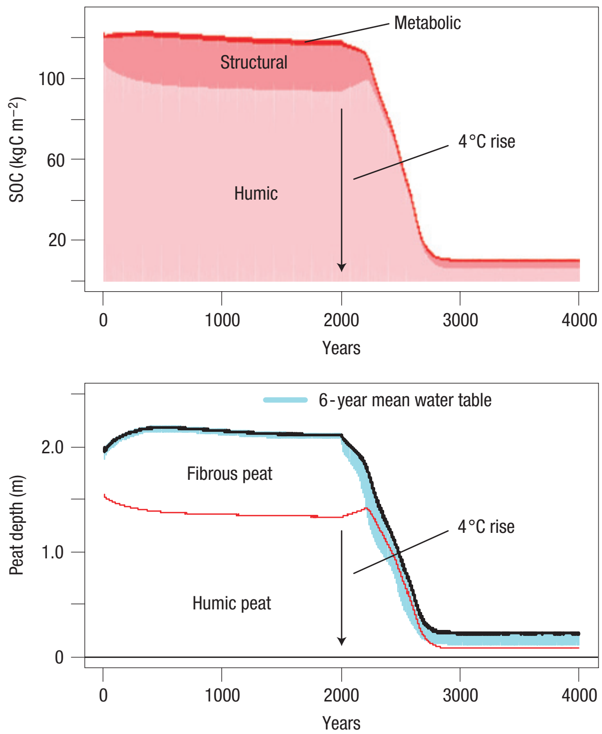

Figure 3 A 4000-year simulation of peat SOC and peat depth at the BOREAS Fen site. a,**b**, Peat SOC (**a**) and peat depth (**b**). Meteorological data for 1994–2005 are used recursively for this long-term simulation. For years 0–2000, the simulated peat column is in dynamic equilibrium under the current climate. A uniform rise of temperature by 4 °C is applied at year 2000, indicated by downward arrows. The black line denotes total peat (fibrous plus humic), and the red line denotes the boundary between fibrous and humic peat.

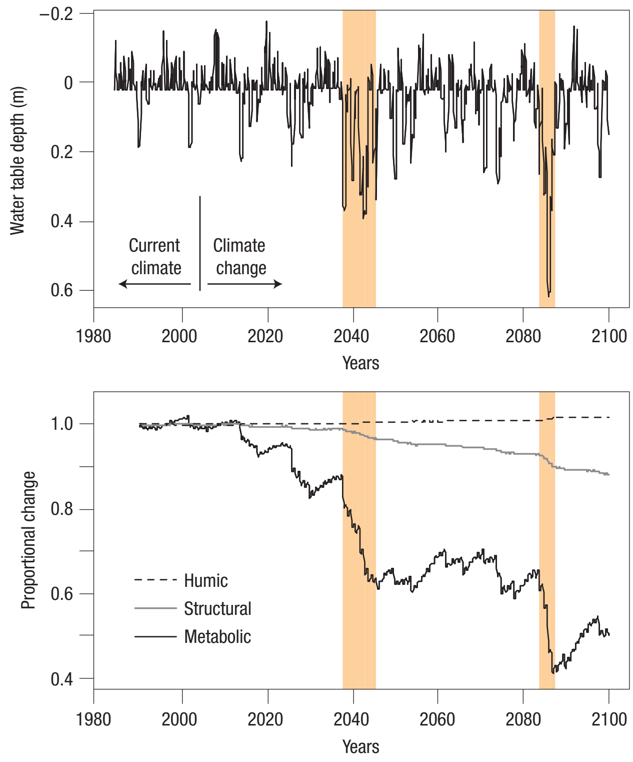

Figure 4 Transient change in the water table at the BOREAS Fen site, 2004–2099. a, Change in water table. **b**, Proportional changes in SOC. Before 2004, the model is in equilibrium under the 1994–2005 climate. Then, temperature and precipitation anomalies projected by HadCM3 with Special Report on Emissions Scenarios A2 are used to force the model in and after year 2004. Shaded areas denote extremely dry periods of 2038–2045 and 2084–2087.

decomposition. The change in temperature triggers this feedback, and the soil water–carbon system is eventually shifted to the new low-SOC regime.

To study the transient behaviour of the system, we disturbed the current equilibrium of the Fen simulation based on the temperature and precipitation anomalies predicted by the general circulation model HadCM3 using scenario A2 of the Special Report on Emissions Scenarios for the period 2004–2099 (ref. 25). Transient responses of SOC to climate change strongly depend on the peat type[2,18] (Fig. 4). The metabolic pool responds quickly to climate change, and the decomposition rate of this pool is controlled by interannual variability in climate. Extended dry periods are indicated during 2038–2045 and 2084–2087 due to climate fluctuations generated by HadCM3 and the hydrological memory of the peat system. The metabolic pool loses more than 20% of SOC during each of these dry periods due to exposure of SOC to aerobic conditions[10]. Although the metabolic SOC is a minor portion of the total SOC, its fast temperature response is the key process of interannual fluctuations in net ecosystem carbon exchange observed in northern peatlands[4].

To single out the physical–biogeochemical interactions, we intentionally omitted ecophysical responses of peatland vegetation to environmental changes. In reality, however, plants will sensitively respond to changes in moisture and temperature regimes, nutrient status, atmospheric CO_2 and peat texture[26,27], and changes in the wetland vegetation community and litter quantity and quality strongly influence peat decomposition and accumulation dynamics[28]. This study emphasizes that the hydrological–biogeochemical feedback inherent to peat has a strong potential to increase climate sensitivities and avoids a study design that might be confounded by two new feedbacks, namely biogeochemical and vegetation dynamics. The CO_2 emissions from the peat collapse predicted by this study could be ameliorated or exacerbated by changes in ecosystem structure and function. Our next research step is to include dynamic vegetation simulated by the ED model framework[15,16].

The transient resistance to peat decomposition observed in the Fen site simulation is due mainly to microbial conversion of labile SOC into more recalcitrant SOC[29]. The massive SOC loss induced by the soil-condition–carbon feedback can be prevented if the temperature rise is reversed within a few hundred years or if a significant increase in precipitation maintains the current levels of the water table[4]. In summary, our modelling approach demonstrates how the mechanistic linkages that exist between the physical and biogeochemical dynamics of peatlands have strong implications for the response of northern peatlands to climate change[30], including a large peat loss due to positive feedbacks in organic soil.

METHODS

Air temperature, wind speed, net radiation and humidity observed for 12 years (1994–2005) at the BOREAS Northern Study Area OBS eddy-covariance tower site every 30 min (<http://www-as.harvard.edu/data >) are repeatedly used to drive the long-term simulations. Missing observations are filled using meteorological data from Thompson Regional Airport (39 km east of the OBS site; <http://climate.weatheroffice.ec.gc.ca >) or by interpolation[4]. Model parameters of biogeochemistry (litter production, lignin-to-nitrogen ratio of litter, and ratio between vascular plants and moss production) and peat biophysics (peat densities, carbon fractions, hydrothermal parameters of fibrous and humic layers) were obtained from previous field and modelling studies[21,24] (see Supplementary Information, Methods). Soil water-table depth and temperature at the OBS site for spring–summer 2003 (Fig. 1) were measured at a subsite in a hollow, where the soil profile contained 2 cm of live *Sphagnum* spp., fibrous dead moss from 2 to 15 cm, partially decayed moss from 15 to 20 cm and humified organic matter from 20 to 35 cm depth.

The dynamics of SOC pool C_i is modelled by $\mathrm{d}C_i/\mathrm{d}t = I_i - TD_l MD_l k_i C_i$, where i is SOC pool type (metabolic, structural, and humic), I_i is the SOC input rate for pool i, l represents the peat layers specified for SOC type i (fibrous layer for metabolic and structural pools, humic layer for humic pool) and k_i is the intrinsic rate of decomposition[2,9,18] (see Supplementary Information, Methods). The temperature and moisture dependence functions TD_l and MD_l are continuously updated based on soil physical conditions. The layer depth Z_l is calculated from SOC in each layer l by $Z_l = C_l/(F_l D_l)$, where F_l is the carbon mass fraction, D_l is the bulk density of soil layer l and $C_{\mathrm{fibrous}} = C_{\mathrm{metabolic}} + C_{\mathrm{structural}}$. Soil hydrology and thermal dynamics are derived from land-surface modelling schemes of ED-RAMS (ref. 16).

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References

1. Gorham, E. Northern peatlands—role in the carbon-cycle and probable responses to climatic warming. *Ecol. Appl.* **1**, 182–195 (1991).
2. Ise, T. & Moorcroft, P. R. The global-scale temperature and moisture dependencies of soil organic carbon decomposition: An analysis using a mechanistic decomposition model. *Biogeochemistry* **80**, 217–231 (2006).
3. Clymo, R. S. The limits to peat bog growth. *Phil. Trans. R. Soc. B.* **303**, 605–654 (1984).
4. Dunn, A. L., Barford, C. C., Wofsy, S. C., Goulden, M. L. & Daube, B. C. A long-term record of carbon exchange in a boreal black spruce forest: Means, responses to interannual variability, and decadal trends. *Glob. Change Biol.* **13**, 577–590 (2007).
5. Anderson, R. L., Foster, D. R. & Motzkin, G. Integrating lateral expansion into models of peatland development in temperate New England. *J. Ecol.* **91**, 68–76 (2003).
6. Hilbert, D. W., Roulet, N. & Moore, T. Modelling and analysis of peatlands as dynamical systems. *J. Ecol.* **88**, 230–242 (2000).
7. Belyea, L. R. & Baird, A. J. Beyond the limits to peat bog growth: Cross-scale feedback in peatland development. *Ecol. Monogr.* **76**, 299–322 (2006).
8. Foster, D. R., Wright, H. E. Jr, Thelaus, M. & King, G. A. Bog development and land-form dynamics in central Sweden and south-eastern Labrador, Canada. *J. Ecol.* **76**, 1164–1185 (1988).
9. Frolking, S. *et al.* Modeling seasonal to annual carbon balance of Mer Bleue Bog, Ontario, Canada. *Glob. Biogeochem. Cycles* **16**, 1030 (2002).
10. Wickland, K. P. & Neff, J. C. Decomposition of soil organic matter from boreal black spruce forest: Environmental and chemical controls. *Biogeochemistry* **87**, 29–47 (2008).
11. Brown, R. J. E. *Proc. 3rd Int. Peat Congr.* 174–181 (National Research Council of Canada, 1970).
12. Zoltai, S. C. Palsas and peat plateaus in central Manitoba and Saskatchewan. *Can. J. Forest Res.* **2**, 291–302 (1972).
13. Fenton, N., Lecomte, N., Legare, S. & Bergeron, Y. Paludification in black spruce (*Picea mariana*) forests of eastern Canada: Potential factors and management implications. *Forest Ecol. Manage.* **213**, 151–159 (2005).
14. Kump, L. R., Kasting, J. F. & Crane, R. G. *The Earth System* 2nd edn (Prentice Hall, 2003).
15. Moorcroft, P. R., Hurtt, G. C. & Pacala, S. W. A method for scaling vegetation dynamics: The ecosystem demography model (ED). *Ecol. Monogr.* **71**, 557–585 (2001).
16. Medvigy, D. M. *The State of the Regional Carbon Cycle: Results from a Constrained Coupled Ecosystem–Atmosphere Model.* PhD dissertation, Harvard Univ. (2006).
17. Pielke, R. A. *et al.* A comprehensive meteorological modeling system—RAMS. *Meteorol. Atmos. Phys.* **49**, 69–91 (1992).
18. Parton, W. J., Schimel, D. S., Cole, C.V. & Ojima, D. S. Analysis of factors controlling soil organic matter levels in Great Plains grasslands. *Soil Sci. Soc. Am. J.* **51**, 1173–1179 (1987).
19. Agren, G. I. & Bosatta, E. Reconciling differences in predictions of temperature response of soil organic matter. *Soil Biol. Biochem.* **34**, 129–132 (2002).
20. Sellers, P. *et al.* The boreal ecosystem–atmosphere study (BOREAS): An overview and early results from the 1994 field year. *Bull. Am. Meteorol. Soc.* **76**, 1549–1577 (1995).
21. Apps, M. J. & Halliwell, D. BOREAS TE-13 biometry reports. Data set. Available online <http://www.daac.ornl.gov> from Oak Ridge National Laboratory Distributed Active Archive Center, Oak Ridge, Tennessee, USA (1999).
22. Carrasco, J. J., Neff, J. C. & Harden, J. W. Modeling physical and biogeochemical controls over carbon accumulation in a boreal forest soil. *J. Geophys. Res. Biogeosci.* **111**, G02004 (2006).
23. Cox, P. M., Betts, R. A., Jones, C. D., Spall, S. A. & Totterdell, I. J. Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model. *Nature* **408**, 184–187 (2000).
24. Bubier, J. L., Crill, P. M., Varner, R. K. & Moore, T. R. BOREAS TGB-01/TGB-03 CH_4 chamber flux data: NSA fen. Data set. Available online <http://www.daac.ornl.gov> from Oak Ridge National Laboratory Distributed Active Archive Center, Oak Ridge, Tennessee, USA (1998).
25. Gordon, C. *et al.* The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. *Clim. Dyn.* **16**, 147–168 (2000).
26. Mack, M. C., Schuur, E. A. G., Bret-Harte, M. S., Shaver, G. R. & Chapin, F. S. Ecosystem carbon storage in arctic tundra reduced by long-term nutrient fertilization. *Nature* **431**, 440–443 (2004).
27. Suzuki, R., Masuda, K. & Dye, D. G. Interannual covariability between actual evapotranspiration and PAL and GIMMS NDVIs of northern Asia. *Remote Sens. Environ.* **106**, 387–398 (2007).
28. Belyea, L. R. & Malmer, N. Carbon sequestration in peatland: Patterns and mechanisms of response to climate change. *Glob. Change Biol.* **10**, 1043–1052 (2004).
29. Giardina, C. P., Ryan, M. G., Hubbard, R. M. & Binkley, D. Tree species and soil textural controls on carbon and nitrogen mineralization rates. *Soil Sci. Soc. Am. J.* **65**, 1272–1279 (2001).
30. Holland, M. M. & Bitz, C. M. Polar amplification of climate change in coupled models. *Clim. Dyn.* **21**, 221–232 (2003).

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Author contributions

T.I. formulated the model framework and conducted the simulations. A.L.D. and S.C.W. designed the field observations, and A.L.D. conducted the fieldwork. The land-surface model ED-RAMS is coded and maintained by P.R.M., and T.I. and P.R.M. wrote the paper. All authors discussed the results and commented on the manuscript.

Author information

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