Is the northern high latitude land-based CO$_2$ sink weakening?


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The Terrestrial Ecosystem Model (TEM) is a monthly, process-based, global-scale ecosystem model that uses spatially referenced information of climate, elevation, soils and vegetation to simulate monthly estimates of terrestrial carbon, nitrogen and water dynamics at different spatial scales. A large number of studies have employed TEM to investigate the major environmental drivers and biophysical processes influencing carbon dynamics in high latitude ecosystems, including boreal forests [Amthor et al., 2001; Potter et al., 2001; Clein et al., 2002; Zhuang et al., 2002], arctic tundra [Clein et al., 2000; McGuire et al. 2000a], and high latitude ecosystems in general [McGuire et al., 2000b; Zhuang et al., 2003, 2004, 2006; Euskirchen et al., 2006]. Recent model versions have incorporated updates that more explicitly address high latitude ecosystem processes, as described by Euskirchen et al. [2006] and Zhuang et al. [2003], including the development of an integrated soil thermal model that considers the effects of freeze-thaw dynamics on gross primary production. In this supplementary section, we provide detail on the updates to model components and process representations related to the simulated fluxes analyzed in this study.

To improve the simulation of terrestrial carbon dynamics including the influence of permafrost on these dynamics in high latitude ecosystems, several modifications from Felzer et al. [2004] have been implemented in version 6.0 of the Terrestrial Ecosystem Model (TEM). These modifications include: 1) changes in the representation of carbon stored in soil organic matter; 2) changes in the potential availability of soil organic matter to decomposition; 3) consideration of dissolved organic carbon (DOC) leaching losses; 4) changes in the simulated influence of temperature on plant maintenance respiration and
decomposition; 5) changes in the estimation of gross primary production (GPP), which represents the uptake of atmospheric carbon dioxide associated with photosynthesis; and 6) the implementation of a dynamic cohort approach to incorporate the influence of disturbances and land use change on the landscape. We conclude with a discussion on model evaluation, where we summarize evidence from other studies on the performance of the model in representing ecosystem processes as site to regional scales.

S1. Representation of organic carbon in soils

In the previous versions, TEM represented the storage of carbon in soil organic matter in two different ways. In the early versions of the model [Raich et al., 1991; McGuire et al., 1992; Melillo et al., 1993], TEM estimated all carbon in soil organic matter as a single pool (Figure S1a) and developed parameters to estimate heterotrophic respiration ($R_H$) based on this aggregated carbon pool ($C_S$). With TEM 4.0 [McGuire et al., 1995], however, it was recognized that much of this soil organic carbon was un-reactive in the context of near-term climate change so TEM was modified to estimate the dynamics of carbon only in “reactive” soil organic matter ($C_{RS}$, Figure S1b). With this modification, TEM 4.0 and later versions of the model would estimate similar values of $R_H$ as the earlier TEM versions for a particular site, but the model parameterizations would be based on the relatively smaller pool size of the reactive soil organic matter rather than all of the organic matter in the soil profile. Because these later versions of TEM did not consider “non-reactive” soil organic carbon ($C_{NS}$), which were assumed to remain constant over potential study periods, estimates of soil organic carbon were less
than similar estimates from the earlier versions of TEM or those derived from field studies.

While the concept of reactive soil organic carbon has been useful for improving estimates of patterns of carbon storage across climate and texture gradients and placing more realistic limits on the response of soil organic matter to future climate change, it has been difficult to compare model results to existing field data and the results of other models. To help facilitate such comparisons and provide a more complete description of carbon dynamics in terrestrial ecosystems, TEM 6.0 now explicitly estimates the amount of “non-reactive” organic carbon in detritus and soils ($C_{NS}$, Figure S1c). As the depth of soil organic matter measurements vary in response to field conditions, the total amount of soil organic carbon assumed to exist at a calibration site is obtained by extrapolating these soil measurements to the rooting depth at the site based on the mean biome-specific distribution of soil organic matter reported by Jobbágy and Jackson [2000]. These estimates are then added to the appropriate estimates of carbon in litter and standing dead to obtain estimates of the total amount of carbon ($C_S$) in detritus and soil organic matter. Estimates of non-reactive carbon in soils at the calibration site ($C_{ns}$) are then determined by subtracting our existing estimate of reactive soil organic carbon ($C_{RS}$) from the corresponding estimate of $C_S$. The ratio of non-reactive soil carbon to total soil organic carbon determined at the calibration site ($C_{ns}:C_{sc}$, Table S1) is then used with the equilibrium estimate of reactive soil organic carbon in a grid cell ($C_{rseq}$) to calculate the amount of non-reactive soil carbon ($C_{ns}$) in a grid cell during model extrapolations as follows:
The same procedure is used to calculate non-reactive soil nitrogen in a grid cell based on the ratio of non-reactive soil nitrogen to total soil organic nitrogen at the calibration site and the equilibrium estimate of reactive soil organic nitrogen in a grid cell. Both non-reactive soil organic carbon and non-reactive soil organic nitrogen are assumed to remain constant throughout a simulation.

[5] In addition to “reactive” and “non-reactive” soil organic carbon pools, the new version of TEM also includes a dissolved organic carbon (DOC) pool (Figure S1c). Incomplete decomposition of soil organic matter is now assumed to produce DOC in the soil in addition to carbon dioxide (Figure S1c). The amount of DOC produced is assumed to be a proportion of the soil organic carbon that decomposes in a month (DOC: Decomposition in Table 2 of McGuire et al. [2010]). This proportion is estimated from DOC export rates [Peterson et al., 1986; MacLean et al., 1999; Petrone, 2005] and the corresponding net primary production [NPP, Raich et al., 1991; McGuire et al., 1992] estimates from intensively studied field sites or, in cases where such data is not available, from vegetation-specific mean DOC export rates [Aitkenhead and McDowell, 2000] and NPP estimates normally used for TEM calibrations [Raich et al., 1991; McGuire et al., 1992]. Similar to the calculation of heterotrophic respiration in earlier TEM versions, decomposition still depends upon the amount of soil organic matter, the C:N ratio of the soil organic matter, air temperature and soil moisture [Raich et al. 1991; McGuire et al. 1997; Tian et al. 1999]. Heterotrophic respiration ($R_h$), however, is now determined by subtracting the amount of DOC produced from the rate of decomposition.
S2. Availability of soil organic carbon to decomposition

[6] In areas underlain by permafrost, the volume of thawed soil or “active layer” varies seasonally as soils freeze and thaw. Thus, the amount of soil organic carbon available for decomposition varies seasonally as organic matter in the frozen portion of the soil profile is protected from decomposition. Furthermore, as the active layer depth increases with global warming, additional soil organic matter within the soil profile may become susceptible to decomposition in the future.

[7] In previous versions of TEM, the potential influence of permafrost on the availability of soil organic matter to decomposition has not been considered. Estimates of heterotrophic respiration in past TEM simulations have assumed that the entire soil organic carbon pool is susceptible to decomposition and the effects of air or soil temperatures and soil moisture within the maximum rooting zone of the dominant vegetation on that decomposition [McGuire et al., 1997; Zhuang et al., 2003; Euskirchen et al., 2006]. In permafrost regions, however, the depth of the active layer is often less than the maximum rooting depth (0.3 to 2.9 m) assumed for these biomes [Vörösmarty et al., 1989] indicating that some of the soil organic carbon stored within the rooting zone is protected from the seasonal or interannual variations in temperature or soil moisture, but may become more susceptible to decomposition with warming-induced degradation of permafrost.

[8] To better account for the influence of permafrost dynamics on terrestrial carbon dynamics, some new features have been incorporated in the new version of TEM. We
have incorporated the soil thermal module (STM) of Zhuang et al. [2003] as modified by Euskirchen et al. [2006] into TEM 6.0 to simulate soil thermal dynamics. The active layer depth determined from the STM influences the relative amount of soil organic carbon available for decomposition within the soil profile and the amount of inorganic nitrogen available for uptake by plants and microbes or leaching to the local water bodies. Because most soil organic matter is located towards the top of the soil profile [Jobbágy and Jackson, 2000], the additional soil organic matter exposed to decomposition at deeper depths in the profile will be less per unit depth than that exposed at shallower depths. Furthermore, the soil organic matter at shallower depths may contain more labile components so that they are more “reactive” to decomposition than organic matter found at deeper depths (Figure S2). To account for these variations, TEM now assumes that the amount of carbon available for decomposition is dependent upon the relative depth of the active layer to the maximum rooting depth as prescribed for an ecosystem. This relative depth changes seasonally each year and may become larger with climate warming.

The hyperbolic functions used for describing the distribution of total soil organic carbon in the soil profile have been developed from the biome-specific horizon information provided in Tables 3 and 4 of Jobbágy and Jackson [2000]:

\[ propSOC = \frac{A(\text{relDepth})}{B + \text{relDepth}} \]

where \( propSOC \) is the proportion of total soil organic carbon down to a particular relative depth (relDepth) to the total amount of total soil organic carbon depth of the rooting zone; and \( A \) and \( B \) are empirical coefficients that vary across biomes (Table S1).
The values of $A$ and $B$ are determined from a nonlinear regression between $propSOC$ and relDepth using the Levenberg-Marquardt algorithm [Bates and Watts, 1988; Seber and Wild, 2003]. Because no information has been available to describe how the lability of soil organic carbon varies with depth, we have assumed that the distribution of reactive soil organic carbon in the soil profile mimics the distribution of total soil organic carbon. Thus, the amount of soil organic carbon available for decomposition in any particular month is determined by multiplying the size of the reactive soil organic pool ($C_{RS}$) by $propSOC$. The proportion of reactive soil organic carbon to total soil organic carbon, however, is assumed to vary across biomes as represented by the proportion of non-reactive carbon to total soil organic carbon ($C_{nsc}:C_{sc}$) listed in Table S1.

The maximum depth of the rooting zone (rootDepth) is not constant, but is assumed to vary with the proportion of silt and clay (PSIPLUSC) in the soil:

$$\text{rootDepth} = Za(\text{PSIPLUSC})^2 + Zb(\text{PSIPLUSC}) + Zc$$

where $Za$, $Zb$ and $Zc$ are vegetation-specific parameters (Table S2) and are derived from regression analyses based on information from Table 1 in Vörösmarty et al. [1989]. In addition, rootDepth is assumed to be greater than or equal to a vegetation-specific minimum rooting depth ($Z_{min}$, Table S2). Thus, the maximum rooting depth can range from 0.8 to 2.9 for forests, 0.55 to 1.7 m for xeric woodlands and shrublands, and 0.3 to 1.2 m for tundra and grasslands (Figure S3).

In our analysis, we track the storage and dynamics of soil organic carbon (SOC) located only within the maximum rooting zone of an ecosystem. The pool of SOC as
defined in the TEM refers to the total amount of organic carbon (but not including inorganic carbon) through the whole soil profile defined by the maximum rooting zone of a particular ecosystem type. This definition should not to be confused with the alternative perspective, which would consider “organic” carbon as only that within the organic layer (and not the mineral layer) of the soil profile. In the TEM, SOC located below the maximum rooting zone [Jobbágy and Jackson, 2000; Tarnocai et al., 2009] is not currently considered by the model and is not assumed to contribute to carbon fluxes back to the atmosphere. Thus, our estimate of approximately 670 Pg C for total SOC stocks in the circumpolar region is much lower than the 1672 Pg C estimated by Tarnocai et al. [2009] and does not account for SOC at deeper depths in deltaic and yedoma soils considered in that study. Our estimate compares somewhat more favorably with the estimate by Tarnocai et al. [2009] of 1024 Pg C when these deeper carbon stocks are not considered. In addition, our estimate of 79.2 Pg C for total SOC stocks in the North American Arctic compares relatively well with the 98.2 Pg C estimated by Ping et al. [2008] for the same region. Ping et al. [2008] estimate SOC stocks to a depth of 1 meter, which corresponds more closely to our maximum rooting depth prescribed for tundra ecosystems, although the area of arctic tundra in North America prescribed in this study (3.79 million km$^2$) is larger than the area (3.040 million km$^2$) considered by Ping et al. [2008].

While total SOC stocks in the circumpolar region are underestimated by TEM, most of this deeper carbon is in perennially frozen soils [Tarnocai et al., 2009] and is currently protected from decomposition to contribute little, if any, carbon fluxes back to the atmosphere. With the limited degradation of the permafrost in our retrospective
analysis, the lack of consideration of carbon stored in permafrost below the maximum rooting depth probably introduced little, if any, bias in our estimates of heterotrophic respiration in this study. As warming continues into the future, however, our current approach may underestimate the influence of permafrost degradation on the availability of carbon to decomposition because the active layer may deepen below the maximum rooting depth.

[13] Yi et al. (2010) and Schaefer et al. (2011) have recently introduced approaches that can account for the effects of a thickening active layer on deeper soil carbon stocks. The approach developed by Yi et al. (2010), which represents an ongoing effort to explicitly couple and represent interactions among soil thermal, hydrological, and biogeochemical processes with depth in TEM (see also Yi et al. 2009a, 2009b), has only been evaluated for wet and dry black spruce forest sites in Alaska, but is currently being evaluated at the regional scale of interior Alaska for a spectrum of vegetation types in different landscape settings.

S3. Leaching of dissolved organic carbon

[14] In previous TEM versions, all losses of carbon from the soil organic carbon pool (C_S or C_RS) have been assumed to occur as carbon dioxide resulting from decomposition (Figure S1a,b). Besides heterotrophic respiration, loss of organic carbon from an ecosystem can occur through DOC leaching losses. In TEM 6.0, the leaching of DOC depends on the runoff of rainwater and snowmelt from the soil profile and the concentration of DOC in the active layer. Thus, no DOC leaching occurs if no runoff is
estimated to occur during a particular month. The leaching rate of DOC (LCHDOC, g C m\(^{-2}\) mo\(^{-1}\)) is determined as follows:

\[
LCHDOC = [\text{DOC}] (\text{rrun} + \text{srun})
\]

where and \([\text{DOC}]\) is the concentration of DOC in soil water (g m\(^{-3}\)); \text{rrun} is the runoff (m mo\(^{-1}\)) associated with rainfall; and \text{srun} is the runoff (m mo\(^{-1}\)) associated with snowmelt.

S4. Influence of temperature on plant maintenance respiration and decomposition

In previous versions of TEM, the rates of both plant maintenance respiration and decomposition depended on the size of the respiring carbon pool and temperature [Raich et al., 1991; McGuire et al., 1992; Tian et al., 1999; Clein et al., 2007]. In addition, decomposition also depended on volumetric soil moisture and the C:N ratio of soil organic matter [McGuire et al. 1997; Tian et al., 1999]. The temperature dependence of both plant maintenance respiration (R\(_m\)) and decomposition, as represented by heterotrophic respiration (R\(_H\)), was modeled as an exponential function:

\[
f_R(T) = e^{rT}
\]

where \(r\) is the instantaneous rate of change in respiration with the change in temperature and \(T\) is the mean monthly air temperature. For decomposition, \(r\) has been assumed to be a constant value of 0.0693, which corresponds to a value of Q\(_{10}\), defined as the rate of increase in respiration with a 10 °C increase in temperature, equal to 2.0. For autotrophic respiration, \(r\) has been assumed to be dependent on temperature [McGuire et al., 1992]. The notion that plant or soil respiration ought to increase exponentially with
temperature, however, has little mechanistic basis, and is often inaccurate when applied over broad temperature ranges [Tjoelker et al., 2001; Lloyd and Taylor, 1994]. It has long been recognized that respiration rates change with temperature less rapidly at higher temperatures, and more rapidly at lower temperatures, than would be predicted by a simple exponential function. While our $Q_{10}$ approach for autotrophic respiration attempts to account for this temperature dependency, further analysis has indicated that the formulation would simulate a maximum autotrophic respiration rate at about 33.6 °C, which is physiologically unrealistic.

In collaboration with Jeff Amthor (currently at the Department of Agriculture, Food & Natural Resources, University of Sydney, Australia), we have developed a new algorithm to improve our description of the effects of temperature on both plant maintenance respiration and decomposition in TEM 6.0:

$$f_R(T) = \left[ (Q_{ref} e^{-\alpha(T-T_{ref})/(T-T_{ref})/10})/(1+e^{(\beta-T)}+e^{(T-\gamma)}) \right]$$

where $f_R(T)$ is the rate of respiration relative to the rate at a reference temperature ($T_{ref}$, 0°C), $Q_{ref}$ is the $Q_{10}$ at $T_{ref}$; $\alpha$ is an empirical parameter; $\beta$ is the temperature at which respiring cells become damaged by low temperatures (respiration drops rapidly at $T < \beta$); and $\gamma$ is the temperature at which respiring cells become damaged by high temperatures (i.e., respiration drops rapidly at $T > \gamma$). For plant maintenance respiration, we have determined parameter values (Table S3) that mimic the changes in leaf respiration with temperature described by Tjoelker et al. [2001] for the temperature range 5-35 °C. With the new algorithm, the maximum autotrophic respiration rate now occurs at 51 °C. For decomposition, we have determined parameter values (Table S3) that mimic the
temperature-respiration relationship described by Lloyd and Taylor [1994] for soil respiration. In TEM 6.0, plant maintenance respiration is still influenced by air temperature while decomposition is influenced by the average soil temperature determined for the top 20 cm of the soil profile.

S5. Estimation of gross primary production

As in TEM 4.3 [Felzer et al. 2004], estimates of gross primary production (GPP) are still dependent upon photosynthetically active radiation (PAR), the capacity of plant canopy to capture sunlight, temperature, moisture, nitrogen availability, and atmospheric concentrations of carbon dioxide and ozone. However, the relationship of GPP to nitrogen availability and temperature has been modified in TEM 6.0 to improve the representation of the effects of nitrogen availability and permafrost on terrestrial carbon dynamics. In addition to changes in the representation of soil organic carbon and nitrogen described above, we also disaggregated the soil available nitrogen pool into five inorganic nitrogen pools in TEM 6.0: a nitrate pool, a reduced nitrogen pool comprised of ammonium and free amino acids, a nitric oxide (NO) pool, a nitrous oxide (N$_2$O) pool and a dinitrogen (N$_2$) pool.

Because TEM has always assumed that the uptake of atmospheric carbon dioxide by plants is limited by nitrogen availability [e.g., Raich et al. 1991; Kicklighter et al. 1999; Sokolov et al., 2008], this change in the representation of soil available nitrogen has influenced how GPP is determined in TEM 6.0. Of the five pools, only nitrate and the “ammonium” pools are assumed to influence plant productivity. The importance of
these two forms of nitrogen to plants, however, is assumed to vary among biomes based on relative plant uptake and nitrogen cycling rates [Aber and Melillo, 2001]. In biomes with low plant nitrogen uptake rates or where nitrification may be limited by soil pH or the availability of ammonium, plants are assumed to uptake mostly reduced forms of nitrogen. In biomes with relatively higher plant nitrogen uptake rates and higher nitrogen cycling rates, plants are assumed to uptake mostly nitrate. Tundra, boreal forests and boreal woodlands are assumed by TEM 6.0 to have an ammonium economy where only ammonium or free amino acids are taken up by the vegetation. In contrast, grasslands, shrublands, xeromorphic forests and woodlands, croplands and pastures are assumed by TEM 6.0 to have a nitrate economy where only nitrate is taken up by the vegetation. In temperate forest ecosystems, the uptake of both nitrate and reduced nitrogen by vegetation are assumed by TEM 6.0 to occur simultaneously. Similar to earlier versions of TEM, the uptake of either nitrate or reduced nitrogen by vegetation depends on the size of the nitrate or reduced nitrogen pool, respectively, volumetric soil moisture, mean monthly air temperature, atmospheric ozone concentrations, the stage of canopy development during regrowth after disturbance, and the carbon-nitrogen status of the vegetation [Pan et al., 1998; Felzer et al., 2004].

In addition to the uptake of ammonium, free amino acids and nitrate, vegetation in TEM 6.0 may also obtain nitrogen from symbiotic nitrogen fixation. Monthly biological nitrogen fixation is determined from monthly evapotranspiration rates after adapting the “Conservative N Fixation” parameterization algorithm of Cleveland et al. [1999] to a monthly time step. Nonsymbiotic nitrogen fixation is determined as a vegetation-specific proportion of biological nitrogen fixation (NNF:BNF in Table S1) based on information
Symbiotic nitrogen fixation is assumed to be the difference between biological nitrogen fixation and nonsymbiotic nitrogen fixation. Inputs from symbiotic nitrogen fixation are added to the vegetation structural nitrogen pool whereas inputs from nonsymbiotic nitrogen fixation are added to the soil organic nitrogen pool.

Because the timing of spring thaw and the duration of the growing season have a large influence on the carbon balance of high latitude ecosystems [Goulden et al., 1997, 1998; Frolking et al., 1996; Frolking, 1997], we also added the freeze-thaw index of Zhuang et al. [2003] to the calculation of GPP to better account for the effects of freeze-thaw dynamics on carbon uptake. The freeze-thaw index, which varies from 0 to 1, represents the proportion of a specific month in which the ground is thawed and is based on simulated soil temperatures at 10 cm depth.

In TEM, the capacity of the plant canopy to capture sunlight varies seasonally with phenology and with canopy development as vegetation regrow from a disturbance. Felzer et al. [2004] represented both of these short-term and long-term dynamics with the \( f(\text{LEAF}) \) multiplier in their representation of GPP. We use exactly the same algorithms in TEM 6.0 as Felzer et al. [2004], but explicitly consider the short-term effects of seasonal phenology \( f(\text{PHENOLOGY}) \) on the capacity of plants to capture sunlight separately from the longer-term effects of canopy development \( f(\text{CANOPY}) \) in our representation of GPP to help clarify how legacy effects of previous disturbance may influence the current plant productivity in an ecosystem.

Similar to earlier versions of TEM [McGuire et al., 1992, 1997; Pan et al. 1998; Sokolov et al., 2008], the calculation of GPP in TEM 6.0 depends on whether or not
nitrogen is limiting GPP. First, monthly GPP is calculated assuming no nitrogen
limitations occur (GPP$\text{C}$):

$$GPP_C = C_{\text{max}} f(PAR) f(\text{PHENOLOGY}) f(\text{CANOPY}) f(T) f(C_a, G_v) f(O3) f(FT)$$

where $C_{\text{max}}$ is the maximum rate of C assimilation; $f(\text{PAR})$ describes the relative effect of
monthly photosynthetically active radiation on GPP [Raich et al., 1991];
$f(\text{PHENOLOGY})$ describes the relative effect of seasonal phenology on GPP based on
monthly leaf area relative to maximum monthly leaf area estimated for the previous 12
months (same as $f(\text{LEAF})$ in Raich et al. [1991]); $f(\text{CANOPY})$ describes the relative
effect of canopy development on GPP by assuming the amount of leaf biomass in the
canopy is related to the total amount of carbon stored in vegetation biomass [Tian et al.,
2003]; $f(T)$ describes the relative effect of monthly air temperature on GPP [Tian et al.,
1999]; $f(C_a, G_v)$ describes the relative effect of atmospheric carbon dioxide ($C_a$) on GPP as
modified by relative canopy conductance ($G_v$) [Raich et al., 1991; Tian et al., 1999];
$f(O3)$ describes the relative effect of ozone on GPP [Felzer et al., 2004]; and $f(FT)$ is the
freeze-thaw index of Zhuang et al. [2003] as modified by Euskirchen et al. [2006].
Unlike $f_R(T)$ described above, $f(T)$ is a parabolic function in which the simulated
optimum carbon uptake is able to adapt/acclimate to local conditions [Tian et al., 1999].

After GPP$_C$ is determined, GPP$_N$ is then calculated based on estimates of
autotrophic respiration($R_A$) and the effects of nitrogen supply on net primary production
(NPP$_N$):

$$GPP_N = NPP_N + R_A$$

$$NPP_N = P_{CM}(\text{NH4UPTAKE} + \text{NO3UPTAKE} + \text{NMOBIL} + \text{SYMNFIX})$$
where $P_{CN}$ is the C:N ratio of newly produced plant tissue [McGuire et al., 1992], NH4UPTAKE is the amount of ammonium and/or free amino acids acquired by plants from the soil, NO3UPTAKE is the amount of nitrate acquired by plants from the soil, NMOBIL is the amount of vegetation labile nitrogen that is mobilized during a particular month [McGuire et al., 1993, 1997; Pan et al., 1998; Tian et al., 1999] and SYMNFIX is the amount of nitrogen fixed by symbiotic nitrogen fixers.

Monthly GPP is then determined by choosing the minimum value of the two estimates of GPP. After GPP is determined, net primary production (NPP) is calculated as the difference between GPP and $R_A$.

### S6. Land cover, land use and disturbance

To enable the evaluation of different disturbance and land-use change events, we have developed a number of spatially-explicit time series data sets to prescribe the timing, area and distribution of historical disturbances and land use change, including wildfire, across the pan-arctic. To represent the influence of disturbance and land-use change on terrestrial carbon dynamics, TEM now uses a dynamic cohort approach. In this approach (described in the main text), disturbances are overlaid over time on an initial set of undisturbed cohorts that collectively sum to the entire land area of the grid cell. The distribution of potential vegetation types was derived from the Global Land Cover Characterization [GLCC; Loveland et al., 2000] version 2 Seasonal Land Cover Regions (SLCR) data set available at 1km (equal-area) resolution for North America and Eurasia. Among the various vegetation data sets available for this region, the GLCC – SLCR map
was chosen for its detail in classification (200+ categories) and its description of vegetation mosaics, thereby allowing flexibility in translating to the upland arctic, boreal and temperate ecosystem types for which the TEM is calibrated. The translated vegetation map, which included 10 upland categories, was aggregated to the 0.5° grid matching the input climate data sets while retaining the area represented by each unique vegetation type within a grid cell as an individual, non-spatial cohort. Wetland cohort areas were assigned to each grid cell based on a 1° x 1° grid cell fraction inundated database [Matthews and Fung, 1987], where wetland area equals the product of fraction inundated and total cell area. The final classification, then, contained 17 categories (in addition to water, ice, and barren) of upland and associated wetland vegetation types. Some ecosystem types, such as polar desert and xeric shrublands, were not allowed to have wetland associates. Although at a relatively coarse resolution (0.5°), the cohortized potential vegetation map was designed to capture upland vegetation community type mosaics and their transitions across spatial gradients, along with spatially-explicit data on wetland extent and type.

S7. Model Evaluation

Throughout the development of TEM, model estimates have been evaluated at site and regional scales in northern high latitude regions in numerous studies. In these evaluations, TEM has generally demonstrated a high degree of skill in representing observed carbon exchange and component processes at monthly and annual timescales. At the site scale, model estimates for northern high latitude ecosystems have been
compared to seasonal and inter-annual field-based estimates for net ecosystem exchange
[Amthor et al., 2001; Clein et al., 2002; Zhuang et al., 2002], soil respiration [Zhuang et
al., 2002], evapotranspiration [Amthor et al., 2001; Zhuang et al., 2002], soil thermal
dynamics [Zhuang et al., 2001, 2002, 2003], and soil moisture dynamics [Zhuang et al.,
2002]. The model results have also been compared with site-specific experimental
responses to warming, nitrogen fertilization, and increased atmospheric CO2
concentration in arctic tundra [Clein et al., 2000]. At the regional scale, model estimates
for northern high latitude ecosystems have been evaluated in the context of analyses
based on atmospheric CO2 data at seasonal [McGuire et al., 2000; Dargaville et al., 2002;
Zhuang et al., 2003], inter-annual [Dargaville et al., 2002], and longer term [Balshi et al.,
2007] scales. The ability of the model to simulate the regrowth of northern high latitude
tree species after disturbance has been evaluated for black spruce [Zhuang et al., 2002]
and aspen [Pan et al., 2002]. The model has also been compared with regional field-
based estimates of net primary production in northern high latitude ecosystems [McGuire
et al. 2002], and remote-sensing estimates in northern high latitudes for gross and net
primary production [Kimball et al., 2007], snow cover [Euskirchen et al., 2006], soil
freeze thaw [Euskirchen et al., 2006], and growing season length [Euskirchen et al.,
2006]. The model estimates of changes in vegetation biomass have been compared with
inventory-based estimates [Zhuang et al., 2003; Balshi et al., 2007]. Runoff estimates of
the model have been evaluated for watersheds across the United States [Gordon et al.,
2004].
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**Supplemental Tables and Figures**

**Table S1.** Parameters used in simulating soil organic carbon dynamics and nonsymbiotic nitrogen fixation.

<table>
<thead>
<tr>
<th>Description of Vegetation</th>
<th>A</th>
<th>B</th>
<th>Cnsc:Csc</th>
<th>NNF:BNF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dwarf shrub arctic tundra/alpine tundra</td>
<td>1.5450</td>
<td>0.4977</td>
<td>0.5*</td>
<td>1.00</td>
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<tr>
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<td>0.4977</td>
<td>0.5*</td>
<td>1.00</td>
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<tr>
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<td>1.2260</td>
<td>0.2398</td>
<td>0.5*</td>
<td>1.00</td>
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<tr>
<td>Boreal needleleaf deciduous forest</td>
<td>1.2260</td>
<td>0.2398</td>
<td>0.5*</td>
<td>1.00</td>
</tr>
<tr>
<td>Boreal broadleaf deciduous forest</td>
<td>1.2260</td>
<td>0.2398</td>
<td>0.5*</td>
<td>1.00</td>
</tr>
<tr>
<td>Temperate needleleaf evergreen forest</td>
<td>1.2132</td>
<td>0.2368</td>
<td>0.5#</td>
<td>0.50</td>
</tr>
<tr>
<td>Temperate broadleaf deciduous forest</td>
<td>1.1539</td>
<td>0.1761</td>
<td>0.5#</td>
<td>0.50</td>
</tr>
<tr>
<td>Xeric woodlands</td>
<td>2.0129</td>
<td>1.0115</td>
<td>0.825</td>
<td>0.35</td>
</tr>
<tr>
<td>Xeric shrubland</td>
<td>1.7556</td>
<td>0.7683</td>
<td>0.75+</td>
<td>0.35</td>
</tr>
<tr>
<td>Grasslands</td>
<td>1.5712</td>
<td>0.5801</td>
<td>0.75+</td>
<td>0.90</td>
</tr>
</tbody>
</table>

*estimate based on Grünzweig et al. [2004]

#estimate based on Gaudinski et al. [2000]

+estimate based on Schlesinger [1997]
Table S2. Parameters used to describe the influence of vegetation type and soil texture on rooting depth (m).

<table>
<thead>
<tr>
<th>Description of Vegetation</th>
<th>Za</th>
<th>Zb</th>
<th>Zc</th>
<th>Zmin (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Forests</td>
<td>0.0000</td>
<td>-2.0875</td>
<td>2.8977</td>
<td>0.8</td>
</tr>
<tr>
<td>Xeric woodlands and shrublands</td>
<td>-6.8820</td>
<td>4.9540</td>
<td>0.7864</td>
<td>0.55</td>
</tr>
<tr>
<td>Tundra and Grasslands</td>
<td>-4.7210</td>
<td>4.1060</td>
<td>0.3003</td>
<td>0.3</td>
</tr>
</tbody>
</table>
**Table S3.** Parameters used to describe the influence of temperature on respiration processes

<table>
<thead>
<tr>
<th>Process</th>
<th>Qref</th>
<th>$\alpha$</th>
<th>$\beta$</th>
<th>$\gamma$</th>
<th>Tref</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plant maintenance respiration</td>
<td>2.07</td>
<td>0.01</td>
<td>-5.0</td>
<td>55.0</td>
<td>25.0</td>
</tr>
<tr>
<td>Decomposition</td>
<td>1.83</td>
<td>0.009</td>
<td>-10.0</td>
<td>100.0</td>
<td>25.0</td>
</tr>
</tbody>
</table>
**Figure S1.** Representation of carbon pools and fluxes in terrestrial ecosystems across various versions of TEM. Carbon pools include vegetation carbon ($C_V$), total soil organic carbon ($C_S$), reactive soil organic carbon ($C_{RS}$), nonreactive soil organic carbon ($C_{NS}$) and dissolved organic carbon (DOC). Carbon fluxes include gross primary productivity (GPP), autotrophic respiration ($R_A$), heterotrophic respiration ($R_H$), production of DOC (DOCPROD) and leaching of DOC from the ecosystem (LCHDOC). Dashed boxes represent carbon pools not explicitly simulated by that version of TEM.
Figure S2. Idealized representation of the distribution of total soil organic carbon (dotted line) and reactive soil organic carbon (solid line) over a soil profile and its relationship to permafrost dynamics when the lability of soil organic matter decreases with depth. Note that as the active layer increases (top layer of permafrost changes from dashed line to dash-dot-dot line), more soil organic carbon becomes exposed to decomposition and that a larger proportion of reactive soil organic carbon in the soil profile may be exposed earlier than indicated by the distribution of total soil organic carbon. In this study, the distribution of reactive soil organic carbon is assumed to be the same as the distribution of total soil organic carbon across the soil profile.
Figure S3. Assumed relationship of maximum rooting depth to soil texture, as represented by proportion silt plus clay, for different vegetation types by TEM.