

Equilibrium responses of soil carbon to climate change: empirical and process-based estimates

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Abstract. We use a new version of the Terrestrial Ecosystem Model (TEM), which has been parameterized to control for reactive soil organic carbon (SOC) across climatic gradients, to evaluate the sensitivity of SOC to a 1°C warming in both empirical and process-based analyses. In the empirical analyses we use the steady state SOC estimates of TEM to derive SOC-response equations that depend on temperature and volumetric soil moisture, and extrapolate them across the terrestrial biosphere at 0.5° spatial resolution. For contemporary climate and atmospheric CO₂, mean annual temperature explains 34.8% of the variance in the natural logarithm of TEM-estimated SOC. Because the inclusion of mean annual volumetric soil moisture in the regression explains an additional 19.6%, a soil moisture term in an equation of SOC response should improve estimates. For a 1°C warming, the globally derived empirical model estimates a terrestrial SOC loss of 22.6 10¹⁵ g (Pg), with 77.9% of the loss in extra-tropical ecosystems. To explore whether loss estimates of SOC are affected by the spatial scale at which the response equations are derived, we derive equations for each of the eighteen ecosystems considered in this study. The sensitivity of terrestrial SOC estimated by summing the losses predicted by each of

the ecosystem empirical models is greater (27.9 Pg per °C) than that estimated by the global empirical model; the 12.2 Pg loss (43.7%) in tropical ecosystems suggests that they may be more sensitive to warming. The global process-based loss of SOC estimated by TEM in response to a 1°C warming (26.3 Pg) is similar to the sum of the ecosystem empirical losses, but the 13.6 Pg loss (51.7%) in extra-tropical ecosystems suggests that they may be slightly less sensitive to warming. For the modelling of SOC responses, these results suggest that soil moisture is useful to incorporate in empirical models of SOC response and that globally derived empirical models may conceal regional sensitivity of SOC to warming. The analyses in this study suggest that the maximum loss of SOC to the atmosphere per °C warming is less than 2% of the terrestrial soil carbon inventory. Because the NPP response to elevated CO₂ has the potential to compensate for this loss, the scenario of warming enhancing soil carbon loss to further enhance warming is unlikely in the absence of land use or changes in vegetation distribution.

Key words. Soil organic carbon, global carbon cycle, climate change, greenhouse effect, model, parameterization, Terrestrial Ecosystem Model.

INTRODUCTION

Terrestrial ecosystems are thought to contain approximately 1500 Pg C (10¹⁵ g carbon) in soil organic matter (Schlesinger, 1977; Post *et al.*, 1982; Eswaran, Van Den Berg & Reich, 1993), which is about twice the carbon in the atmosphere (Watson *et al.*, 1990). Carbon is transferred from soils to the atmosphere predominantly in the form of CO₂ through the process of decomposition; this flux is approximately 50 Pg C annually (Watson *et al.*, 1990). A dominant feature of climate change predicted by general circulation models (GCMs) for a doubled CO₂ atmosphere is an increase in mean surface temperature of the globe (Mitchell *et al.*, 1990). Because decomposition is a temperature-sensitive process (Peterjohn *et al.*, 1993, 1994; Kicklighter *et al.*, 1994), elevated temperature has the

potential to increase the flux of CO₂ from soils to the atmosphere. However, if plant production increases in response to climate change (see Melillo *et al.*, 1993), then the removal of carbon from the atmosphere may be enhanced. Thus, climate change has the potential to either accelerate or slow the build-up of atmospheric CO₂ depending on how it influences both decomposition and plant production. Clearly, assessments of the influence of climate change on the atmospheric concentration of CO₂ require models that can integrate the effects of climate change on carbon storage in the terrestrial biosphere.

In this study we use a new version of the Terrestrial Ecosystem Model (TEM), which is a process-based model that has previously been used to assess the effects of climate change and elevated CO₂ on soil organic carbon (SOC; Melillo *et al.*, 1995; McGuire, Kicklighter &

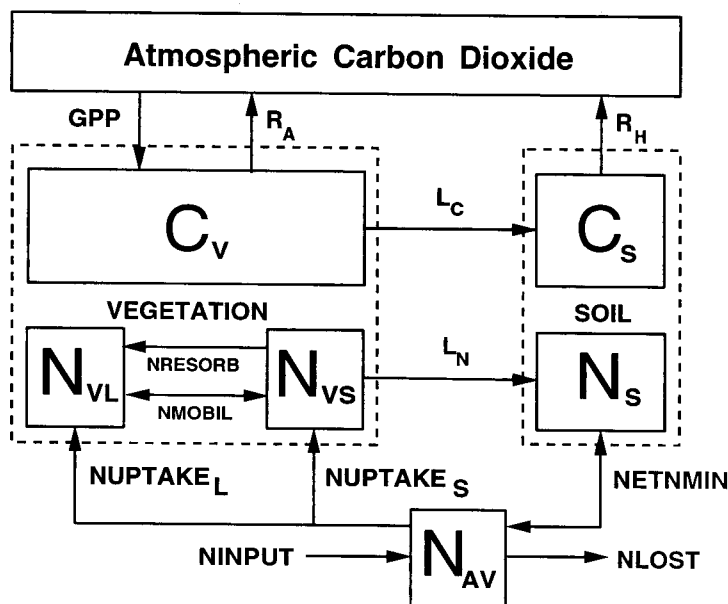


FIG. 1. The Terrestrial Ecosystem Model (TEM). The state variables are: carbon in the vegetation (C_V); structural nitrogen in the vegetation (N_{VS}); labile nitrogen in the vegetation (N_{VL}); organic carbon in soils and detritus (C_S); organic nitrogen in soils and detritus (N_S); and available soil inorganic nitrogen (N_{AV}). Arrows show carbon and nitrogen fluxes; GPP , gross primary productivity; R_A , autotrophic respiration; R_H , heterotrophic respiration; L_C , litterfall carbon; L_N , litterfall nitrogen; $NUPTAKE_S$, nitrogen uptake into the structural nitrogen pool of the vegetation; $NUPTAKE_L$, nitrogen uptake into the labile nitrogen pool of the vegetation; $NRESORB$, nitrogen resorption from dying tissue into the labile nitrogen pool of the vegetation; $NMOBIL$, nitrogen mobilized between the structural and labile nitrogen pools of the vegetation; $NETNMIN$, net nitrogen mineralization of soil organic nitrogen; $NINPUT$, nitrogen inputs from outside the ecosystem; and $NLOST$, nitrogen losses from the ecosystem.

Melillo, 1995), to evaluate the sensitivity of SOC to a 1°C warming in both empirical and process-based analyses. In the empirical analyses, we evaluate the sensitivity of SOC to warming by extrapolating across the terrestrial biosphere empirical models of SOC response to warming; the empirical models are based on the steady state process-based SOC estimates of TEM for contemporary climate and atmospheric CO_2 and include the effects of temperature and soil moisture. In the process-based analyses we estimate SOC change by the difference between the pool sizes of SOC estimated by TEM for contemporary climate and the warming scenario. We also document the influence of elevated CO_2 on SOC estimates of TEM. Finally, we examine the potential implications of SOC response for the global carbon cycle.

MODEL DESCRIPTION AND METHODS

The TEM uses spatially referenced information on climate, soils, and vegetation to make monthly estimates of important carbon and nitrogen fluxes and pool sizes (Fig. 1). The first two versions of TEM were used to examine patterns of net primary production (NPP) in South America (Raich *et al.*, 1991) and North America (McGuire *et al.*, 1992). The third version of TEM was used to examine the response of net primary production (NPP) to elevated temperature and carbon dioxide for temperate forests (McGuire *et al.*, 1993) and to climate change predicted by general circulation models (GCMs) for the terrestrial biosphere (Melillo *et al.*,

1993). The carbon storage predictions of the third version were also evaluated for global terrestrial ecosystems (Melillo *et al.*, 1995) and for grasslands and conifer forests (McGuire *et al.*, 1995). In this study we use a new version of TEM (version 4.0), which we developed to improve patterns of carbon storage along gradients of temperature, moisture and soil texture.

In this paper we describe the modifications to the model that define version 4.0 of TEM, but the details of how the changes were implemented will appear elsewhere. Some of the modifications to TEM alter the effects of temperature, moisture and soil texture on model operation. The temperature modifications influence the equation for gross primary production (GPP). To allow local temperature adaptation/acclimation for carbon uptake, the optimum temperature of GPP is defined by the mean monthly temperature that corresponds to the month of maximum leaf area. Also, the minimum and maximum temperature constraints for GPP are set so that they reflect the growing-season limits of the vegetation type.

The moisture modifications affect nutrient diffusion in the soil, the carbon to nitrogen ratio of vegetation in grasslands and the decomposition of soil organic matter. The diffusion component of nitrogen uptake by plants and microbes has been modified to depend on volumetric soil moisture instead of percentage saturation, which is more consistent with the treatment of nutrient diffusion in Papendick & Campbell (1981). The half-saturation constants for plant and microbial uptake were also modified to depend on

nitrogen concentration in the soil solution instead of area-defined inorganic nitrogen. The vegetation carbon to nitrogen ratio in grasslands now depends on water balance so that lower ratios occur in drier environments, which is similar to the ratio calculations of Parton *et al.* (1987). The effects of moisture on decomposition were modified so that they depend on volumetric soil moisture instead of percentage saturation and the direct effects of soil texture in the moisture–decomposition relationship have been removed. However, soil texture indirectly affects the relationship by influencing volumetric soil moisture.

In previous versions of the model, soil texture was handled as a categorical variable with five classes (sand, sandloam, loam, clayloam and clay). In this version we treat soil texture as a continuous variable based on the proportion of silt plus clay (PSIPLUSC). Relationships have been incorporated into TEM so that rooting depth, porosity, field capacity and wilting point all depend on the PSIPLUSC. In this version of TEM, the parameters for carbon and nitrogen uptake capacity by vegetation are linear functions of PSIPLUSC so that they represent how potential leaf area and potential fine root surface area varies with soil texture. The parameters for decomposition and immobilization capacity by microbes are inversely related to PSIPLUSC so that they represent the physical protection of soil organic matter in fine-textured soils.

A number of miscellaneous modifications are also implemented in this version of TEM. The mean state of decay of the most recent 6 years of litter inputs, which is D in the immobilization equation, has been modified so that it indicates the mean state of decay of SOC for which the proportion of mass remaining from the original litter input is more than 20%. In long-term studies of mass loss from litter cohorts, the dynamics of decomposition appear to qualitatively change as detritus passes from a fast to a slow pool (Aber, Melillo & McLaugherty, 1990); we chose 20% as the transition between a fast and a slow pool based on the data of Aber *et al.* (1990). To implement better control of the carbon to nitrogen ratio of soil organic matter, which is defined by the parameter S_{CN} , the equation for net nitrogen mineralization has been modified to depend on S_{CN} . The calculation for monthly net irradiance, i.e. the shortwave irradiance at the top of the canopy, has been modified to depend on the relationship of Black, Bonython & Prescott (1954) instead of Black (1956). Finally, in this version of TEM we restrict the labile nitrogen pool of the vegetation to be less than 20% of the structural nitrogen pool of the vegetation. This prevents the labile pool from getting unrealistically large, a problem we observed for only a handful of sites in our global extrapolations of TEM.

The data sets used in this study to run TEM (vegetation (Melillo *et al.*, 1993), elevation (NCAR/Navy, 1984), soil texture as percentage of silt plus clay (FAO/CSRC/MBL, undated), mean monthly temperature (Legates & Willmott, 1990a), monthly precipitation (Legates & Willmott, 1990b), and mean monthly percentage cloudiness (Hahn/CSRC/MBL, undated)) are gridded at a spatial resolution of 0.5° (latitude \times longitude). Hydrological inputs for TEM are determined by a water balance model (Vorosmarty *et al.*, 1989) that uses these variables. The application of TEM

to a grid cell requires the use of the monthly climatic and hydrological data and the soil- and vegetation-specific parameters appropriate to the grid cell. The soil-specific parameters define rooting depth, porosity, field capacity and wilting point. Although many of the vegetation-specific parameters are defined from published information (Raich *et al.*, 1991; McGuire *et al.*, 1992), some are determined by calibrating the model to the fluxes and pool sizes of an intensively studied field site that is representative of the vegetation type. Four vegetation types are not parameterized for TEM 4.0 because we consider them to be mosaics of other vegetation types: boreal woodland, temperate savanna, temperate mixed forest and tropical savanna. Boreal woodland is defined to consist of moist tundra and boreal forest, temperate savanna of tall grassland and temperate deciduous forest, temperate mixed forest of temperate deciduous forest and temperate conifer forest and tropical savanna of tall grassland and tropical deciduous forest. The model makes estimates for mosaic grid cells by equally weighting estimates made with each of the appropriate non-mosaic calibrations.

Table 1 indicates the calibration site of each of the non-mosaic vegetation types. Note that although there are fourteen non-mosaic vegetation types, there are only ten parameterizations because some of the vegetation types use the same parameterization. The values of annual NPP, NUPTAKE, C_V , vegetation carbon to nitrogen ratio (V_{CN}), C_S , and soil carbon to nitrogen ratio (S_{CN}) used to calibrate the model for the soil texture of the calibration site are documented in Table 1. Except for C_S , the studies used to determine these values are documented in McGuire *et al.* (1992) for most of the calibration sites.

The pool C_S , which we equate with SOC in the context of TEM, includes dead organic carbon in detritus and soils. The processes affecting C_S in TEM are appropriate for upland aerobic soils; anaerobic soil processes are not represented in the model. For grasslands, we used the equation of Burke *et al.* (1989) for non-cultivated soils to determine the 20 cm C_S appropriate for the soil texture of the Pawnee Grassland calibration site. Although the Burke equation technically is designed to estimate non-detrital dead organic carbon in soils of U.S. grasslands, our application of it for sandloam at the Pawnee site indicates that it may overestimate total dead organic carbon in detritus and soils by almost 20% when compared with information in Table 1 of Clark (1977). Because detritus is a minor component of dead organic carbon in short grasslands and there is substantial spatial variability in soil carbon within a 0.5° grid cell (Burke *et al.*, 1990, 1991), the use of the Burke equation to estimate total dead organic carbon in U.S. grasslands is well within the range of uncertainty and provides a spatially explicit comparison for TEM estimates of C_S in U.S. grasslands.

To control for soil carbon quality across climatic gradients, we determined values of C_S for other calibration sites so that transitions in SOC are smooth at ecosystem boundaries. This was accomplished by calibrating the vegetation-specific parameters of the model to several values of C_S , extrapolating each calibration near the ecosystem boundary of interest, and choosing the C_S that resulted in

