An Efficient Numerical Terrestrial Scheme (ENTS) for Earth System Modelling

M. S. Williamson\textsuperscript{a}, T. M. Lenton\textsuperscript{a,*},
J. G. Shepherd\textsuperscript{b}, N. R. Edwards\textsuperscript{c}

\textsuperscript{a}Tyndall Centre, UK and School of Environmental Sciences, University of East Anglia, Norwich NR4 7TJ, UK
\textsuperscript{b}Tyndall Centre, UK and School of Ocean and Earth Science, National Oceanography Centre, Southampton SO14 3ZH, UK
\textsuperscript{c}Earth Sciences, Open University, Milton Keynes MK7 6AA, UK

Abstract

We present a minimal spatial model of vegetation carbon, soil carbon and soil water storage and the exchange of energy, water and carbon with the atmosphere. The Efficient Numerical Terrestrial Scheme (ENTS) is designed for long time period simulations and large ensemble studies in Earth system models of intermediate complexity (EMICs). ENTS includes new parameterisations of vegetation fractional cover and roughness length as functions of vegetation carbon, and a relationship between soil carbon storage and soil water holding capacity. We make and justify the approximation that when the solar forcing is a diurnal average, as in our EMIC, the land radiation balance equilibrates with the atmosphere within a few days. This allows us to solve directly for equilibrium land temperature, making ENTS very computationally efficient and avoiding problems of numerical instability that beset many land surface schemes. We tune the carbon cycle parameters towards observed values of global carbon storage in vegetation and soil and estimated global fluxes of net photosynthesis, vegetation respiration, leaf litter and soil respiration. When the model is forced with long term monthly mean fields of NCEP reanalysis climate data, we find ENTS yields broadly accurate patterns of vegetation and soil carbon storage, vegetation fraction, surface albedo, land temperature and evaporation.

Key words: Model; land surface; carbon cycle; Earth system; vegetation; soil; photosynthesis; respiration.

* Corresponding author. Tel.: +44 1603 591414; fax.: +44 1603 591327.

Email addresses: m.williamson@uea.ac.uk (M. S. Williamson),
t.lenton@uea.ac.uk (T. M. Lenton), jgs@noc.soton.ac.uk (J. G. Shepherd),
n.r.edwards@open.ac.uk (N. R. Edwards).

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1 Introduction

Minimal models of the interactions between vegetation, climate, and the carbon cycle have previously been developed as globally-averaged components of zero-dimensional box models (Svirezhev and von Bloh, 1997; Adams, 2003). Here we attempt to produce such a minimal model of energy, water and carbon storage on land and exchange with the atmosphere, for use in a spatial context. By ‘minimal’ model we mean the number and descriptions of processes represented are not over detailed for the context in which they are used.

Land surface schemes used in global climate models have a long history (Sellers et al., 1997). The first generation of schemes (e.g. Manabe (1969)) were based on a single soil water bucket with no vegetation canopy and bulk transfer formulae. The second generation of models (e.g. BATS (Dickinson et al., 1986) and SiB (Sellers et al., 1986)) recognized vegetation effects but were restricted to treating the fluxes of energy and water. The third generation of models (e.g. MOSES (Cox et al., 1999; Essery et al., 2001)) improved the representation of photosynthesis and stomatal conductance and coupled the fluxes of carbon to those of energy and water. The inclusion of carbon fluxes in turn allowed the coupling of dynamic global vegetation models (DGVMs, e.g. TRIFFID (Cox, 1998), VECODE (Brovkin et al., 1997)) that capture changes in vegetation and soil carbon storage.

Global models have also diversified with the rise of Earth system models of intermediate complexity (EMICs, Claussen et al. (2002)) and an ongoing increase in the resolution and complexity of general circulation models (GCMs). GCMs represent the Earth system with high spatial and/or temporal resolution and use low level process descriptions making them potentially the most credible predictive tools. The disadvantage of GCMs is the computational cost of running the model through time. Deciphering and understanding the processes that give a particular result in a complex model can also be very challenging. EMICs generally have lower spatial resolution and higher level parameterisations. Their advantages are that they can incorporate a larger number of components of the Earth system, e.g. ocean sediments or economies, they may be integrated for longer periods of time for a given computational resource, and they are more amenable to analysis than GCMs; figuring out what is happening in a simulation is made less daunting by the reduction in complexity of the process descriptions. EMICs retain more in common with the real Earth system than box models, and thus span the gap in model complexity between box models and GCMs.

The Efficient Numerical Terrestrial Scheme (ENTS) is the result of an attempt to construct a minimal, realistic land surface and terrestrial carbon model with high computational efficiency. ENTS was designed to be coupled
into EMICs, in particular the Grid ENabled Integrated Earth system model (GENIE) framework, which is used for long integrations (of $10^3$-$10^6$ yrs), to study past Earth system dynamics and make long-term future projections. Our key requirements were to capture the broad spatial patterns of global land carbon storage in vegetation and soil and the fluxes of moisture, heat and carbon dioxide between the land and the atmosphere. To achieve this we have simplified the relevant equations by introducing some new parameterisations and approximations.

In common with other land surface schemes, the same equations are applied to each land grid cell, and in principle ENTS could be applied on any grid. However, it was developed at the relatively coarse baseline surface resolution of GENIE (as used by Edwards and Marsh (2005)): an equal area, longitude-sine(latitude) global spatial grid of 36x36 cells, which has 362 land surface grid cells. Because of the simplicity of the model, tuning it to available data is relatively straightforward. We describe the coupling of ENTS with a fast 3D ocean model, sea-ice, marine biogeochemistry and a simple 2D energy moisture balance atmosphere in a separate paper (Lenton et al., 2006). The resulting EMIC, GENIE-1, uses an ocean time step of 3.65 days, resolving the seasonal cycle but not the diurnal cycle.

In many Earth system models, the temporal evolution of the land temperature is solved by a finite step numerical integration algorithm that calculates the variable at each time step. Because of the small size of the land heat reservoir and the comparatively large size of the heat fluxes, the time step must be made small to avoid numerical instabilities. This improves the accuracy of the calculation but increases the number of operations required for the simulation of a given length of time and so uses greater computational resources. The skin temperature of the land is a result of radiative fluxes at the land-air interface and of the small conduction of heat creating a vertical temperature gradient. Previous work has captured this vertical structure by creating a model with multiple soil layers, e.g. Cox et al. (1999) who use 4 layers. However, we take the simpler and faster approach of using a single soil layer, e.g. Meissner et al. (2003). We show that when used in a model such as GENIE-1 that does not resolve the diurnal cycle, ENTS can be solved directly for the equilibrium land temperature, thus greatly increasing its computational efficiency.

ENTS represents a hybrid of a simple bucket model (as in first generation land surface schemes) with an explicit but simplified carbon cycle (inspired by third generation schemes). It combines carbon fluxes and stores in one model, rather than having a land surface scheme generating the carbon fluxes and a separate DGVM handling the carbon stores (as in e.g. MOSES-TRIFFID). This allows us to include new connections between vegetation carbon and surface roughness length and between soil carbon content and soil water holding capacity. ENTS does not distinguish between different types of vegetation or
treat the competition of plant functional types, hence it differs from DGVMs. In the GENIE framework, a DGVM (TRIFFID) is available together with a more sophisticated and computationally demanding land surface scheme derived from MOSES (Cox et al., 1999). In terms of vegetation types, ENTS is one step simpler than the VECODE model (Brovkin et al., 1997) widely used in EMICs, which distinguishes trees and grasses. However, the simplicity of ENTS has the advantage that it could be applied to intervals in geologic time when the types of extant vegetation are largely unknown. The single generic vegetation type in ENTS is best thought of as trees, since trees dominate the data used for model calibration. Trees have been abundant for over 350 million years, whereas grasses are relatively recent.

The paper is organized as follows: In section 2 we describe the new model equations. Section 3 describes the model tuning. We compare the model results with data in section 4. Section 5 offers some conclusions. The Appendix gives a more rigorous defence of our method of calculating the land energy balance.

2 Model description

2.1 Land radiation balance

The land-atmosphere radiation balance is sketched in Figure A.1. The land energy balance equations are inspired by the energy balance model that exists in the EMBM of Weaver et al. (2001) and Edwards and Marsh (2005). The solar insolation forcing in GENIE is a diurnal average with a seasonal oscillation. With this forcing and a single soil layer we make the approximation that the land temperature reaches equilibrium more rapidly than the frequency of each call to the land (3.65 days). This approximation is formally justified in the Appendix. We therefore solve for $T_l$ in the following energy balance in equilibrium condition:

$$(1 - \alpha_{atm})(1 - C_A)(1 - \alpha_s)Q_{SW} = Q_{LH} + Q_{LW} + Q_{SH} \quad (1)$$

where $Q_{SW}$ is the incoming short wave solar radiation at the top of the atmosphere, $\alpha_{atm}$ is the atmospheric albedo, $C_A = 0.3$ is a partitioning parameter determining the quantity of radiation absorbed by dust and moisture in the atmosphere, and $\alpha_s$ is the land albedo (see section 2.4). The heat flux terms on the right hand side of equation (1) are given by the following equations and have dimensions of power per unit area (units of W m$^{-2}$).
Latent heat, $Q_{LH}$ is given by

$$Q_{LH} = \rho_0 L_v E$$

where $\rho_0$ is a reference density of water, 1000 kg m$^{-3}$ and $L_v = 2.50 \times 10^6$ J kg$^{-1}$ is the latent heat of vaporization. $E$ is the evaporation.

Net longwave radiation flux between the land and the atmosphere, $Q_{LW}$ is given by

$$Q_{LW} = \varepsilon_l \sigma T_l^4 - \varepsilon_a \sigma T_a^4$$

where $\varepsilon_l = 0.94$ and $\varepsilon_a = 0.85$ are the emissivity of land and atmosphere respectively. $\sigma = 5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$ is the Stefan-Boltzmann constant and $T_a$ is the air temperature.

Sensible heat, $Q_{SH}$ is given by

$$Q_{SH} = \rho_a C_H c_{pa} U (T_l - T_a)$$

where $\rho_a = 1.25$ kg m$^{-3}$ is a constant surface air density, $c_{pa} = 1004$ J kg$^{-1}$ K$^{-1}$ is the specific heat capacity of air and $U$ is the wind speed in m s$^{-1}$. $C_H$ is a transfer coefficient for heat and is given by

$$C_H = \left[ \frac{1}{\kappa} \ln\left( \frac{z_r}{z_0} \right) \right]^{-2}$$

Where $\kappa = 0.41$ is the Von Karman constant, $z_r = 10$ m is a reference height, the same height at which $U$ and $T_a$ are taken, and $z_0$ is the roughness length of the surface.

A new simple parameterisation for surface roughness length is introduced based on regressing global data sets of vegetation carbon, $C_v$, from Olson et al. (1985) and roughness length, $z_0$, derived from NCEP data (Figure A.2). Roughness length was derived from NCEP long term monthly mean fields of potential evaporation, surface air temperature, wind speed, relative humidity and skin temperature, by assuming potential evaporation takes the form of equation (8) with $\beta = 1$. Roughness length, $z_0$, is found to depend linearly on vegetation carbon, $C_v$:

$$z_0 = \min\{0.001 \text{ m}, k_z C_v \}$$

$k_z$ is given in Table 1. This new parameterisation of roughness length is simpler and has less degrees of freedom than previous land surface schemes, many of which use different roughness lengths for different plant functional types (PFTs). Cox (1998) parameterise roughness length in terms of plant height which is in turn a function of the amount of vegetation carbon present in the woody parts of the plant. This height effect is captured to first order
by our parameterisation, but any non-linearity is ignored. Although we don’t distinguish PFTs, our linear dependence of roughness length on biomass also captures to first order the observation that low biomass grasses have a much shorter roughness length than high biomass trees.

2.2 Land hydrology

Each land point is modelled as having a water bucket with a field capacity of $W_s^*$ and a moisture content of $W_s$ on the current timestep. Both have dimensions of metres. Water is added to the land by precipitation, $P_w$ (m s$^{-1}$) and removed by total evaporation $E$ and runoff $R$:

$$\frac{dW_s}{dt} = P_w - E - R$$  \hspace{1cm} (7)

Runoff occurs only when the ‘bucket’ in each grid box exceeds its field capacity and hence it can be regarded as surface runoff. Other models, e.g. Meissner et al. (2003), use a ‘leaky bucket’ approach and hence can have an additional groundwater runoff term proportional to the soil saturation.

Our formulation for evaporation, $E$, is identical to the bulk formulation approach used in Weaver et al. (2001) and Edwards and Marsh (2005). We calculate the transfer coefficient using a parameterisation for roughness length, $z_0$, given in equation (6).

If the grid box is below field capacity (i.e. $W_s < W_s^*$) then evaporation has a multiplier $\beta$, which reduces the amount of evaporation when the soil is not saturated and water is not freely available to evaporate. The $\beta$ multiplier attempts to model the land surface being less likely to give up water the drier it becomes. This gives $E$ as

$$E = \beta \frac{\rho_a C_W U}{\rho_0} (q_s(T_l) - q_a)$$  \hspace{1cm} (8)

$C_W$ is the transfer coefficient for moisture and we make the reasonable approximation that $C_W = C_H$. The $\beta$ function is formulated to give a value between 0 and 1 depending on the soil saturation. If $\beta = 1$ then $E$ just reverts to evaporation for a surface with freely available water. If $W_s < W_s^*$ then

$$\beta = \left( \frac{W_s}{W_s^*} \right)^4$$  \hspace{1cm} (9)

and $\beta = 1$ if $W_s \geq W_s^*$.

$q_s(T_l)$ is the saturation specific humidity for land and $q_a$ is the atmospheric sur-
face specific humidity (both dimensionless). The saturation specific humidity is based on a parameterisation given by Bolton (1980):

$$q_s(T) = c_1 e^{c_4 T/c_5}$$  \hspace{1cm} (10)

Where the constants are $c_1 = 0.0038$, $c_4 = 17.67$ and $c_5 = 243.5^\circ C$.

When the soil water exceeds field capacity the excess runs off into the ocean, to a location determined by a runoff map as in Edwards and Marsh (2005). If during a timestep, length $\delta t$, $W_s > W_s^*$ then the runoff, $R$ (units m s$^{-1}$) will be

$$R = \frac{1}{\delta t} (W_s - W_s^*)$$  \hspace{1cm} (11)

Where $R$ is always $\geq 0$.

Finally, the soil field capacity has a linear dependence on soil carbon, $C_s$:

$$W_s^* = \min\{k_8, k_9 + k_{10} C_s\}$$  \hspace{1cm} (12)

This novel formulation, inspired by Adams (2003), attempts to capture the difference between the water holding capacity of a desert (low to zero soil carbon) and a peat bog (high soil carbon).

2.3 Land carbon cycle

The land carbon cycle is based on a similar structure to Lenton (2000) and Adams (2003) with revised functions for water stress, leaf litter fall, and the temperature responses of photosynthesis, plant respiration, and soil respiration.

In our model the land can store carbon either in vegetation, $C_v$ (representing living biomass), or in soil, $C_s$ (which includes litter and soil organic carbon but not inorganic carbonates). Vegetation removes carbon from the atmosphere through photosynthesis, $P$, and adds carbon to the atmosphere through plant respiration, $R_v$. Vegetation also loses carbon to the soil through leaf litter, $L$, and soil loses carbon to the atmosphere via soil respiration, $R_s$.

The governing equations are

$$\frac{dC_v}{dt} = P - R_v - L$$  \hspace{1cm} (13)

$$\frac{dC_s}{dt} = L - R_s$$  \hspace{1cm} (14)

Where $C_v$ and $C_s$ have dimensions of carbon per unit area and the carbon...
flux terms $P$, $R_v$, $L$ and $R_s$ have dimensions of carbon per unit area per unit time.

Net photosynthesis, $P$, (i.e. gross photosynthesis minus photo-respiration) is given by

$$P = k_{18} f_1(CO_2) f_2(W_s) f_3(T_a) f_v$$  \hspace{1cm} (15)$$

This is a product of four functions representing the effects of carbon dioxide $f_1(CO_2)$, water stress $f_2(W_s)$, air temperature $f_3(T_a)$, and a saturating dependence on biomass $f_v$. The constant $k_{18}$ is a baseline leaf net photosynthesis rate.

The $CO_2$ response of photosynthesis follows a hyperbola above a compensation point, as in Lenton (2000). When $pCO_2 > k_{13}$ (the compensation point):

$$f_1(CO_2) = \frac{1}{k_{19}} \frac{pCO_2 - k_{13}}{pCO_2 - k_{13} + k_{14}}$$  \hspace{1cm} (16)$$

otherwise $f_1(CO_2) = 0$. $k_{19} = (278 - k_{13})/(278 - k_{13} + k_{14})$ normalizes the $CO_2$ response. Values of $k_{13}$ and $k_{14}$ (Table 1) are taken from Lenton (2000). Our use of a saturating Michaelis-Menten type response is more accurate than the $\beta$-factor logarithmic response used in some other models, e.g. VECODE (Brovkin et al., 1997). However, $CO_2$ is not varied in the present study.

The soil water response of photosynthesis is linear between limiting values. If $\frac{1}{2} W_s^* \leq W_s \leq \frac{3}{4} W_s^*$ then

$$f_2(W_s) = \frac{4W_s}{W_s^*} - 2$$  \hspace{1cm} (17)$$

This increases from 0 at $W_s/2$ to a maximum of 1 at $3W_s/4$. Other models use qualitatively similar linear or saturating responses (Adams et al., 2004).

The temperature response of photosynthesis is based on that of the maximum carboxylation rate of Rubisco, with an additional cut-off term at sub-zero temperatures (following MOSES/TRIFFID, Cox et al. (1998)). However, we combine two temperature response functions in order to capture the response of both high and low latitude vegetation types.

$$f_3(T_a) = f_{3a}(T_a) + f_{3b}(T_a)$$  \hspace{1cm} (18)$$

The two functions making up $f_3(T_a)$ are given by

$$f_{3a}(T_a) = \frac{2.0^{0.1(T_a - T_{ref})}}{(1 + e^{0.3(T_a - k_{11a})})(1 + e^{-0.3(T_a - k_{12a})})}$$  \hspace{1cm} (19)$$

$$f_{3b}(T_a) = \frac{2.0^{0.1(T_a - T_{ref})}}{(1 + e^{0.6(T_a - k_{11b})})(1 + e^{-0.3(T_a - k_{12b})})}$$  \hspace{1cm} (20)$$
Where $T_{\text{ref}} = 298.15\text{K}$. The resulting function is illustrated in Figure A.3.

The two peaks are a novel feature compared to other models (Adams et al., 2004), although some have different temperature responses for C3 and C4 photosynthesis, e.g. Foley et al. (1996). The twin peaks were introduced having analysed vegetation carbon storage data (Olson et al., 1985) as a function of latitude, which shows a main peak at the equator and a secondary peak at high latitudes, representing the tropical and boreal forest zones, respectively. The twin peaked function enables a good match to vegetation and soil carbon storage data (see Section 4), without having to include different high and low latitude vegetation types. We note that the photosynthetic response of plants has considerable capacity to acclimate to prevailing temperatures (Medlyn et al., 2002) and that some boreal forest trees can approach the low optimal temperature for growth that is the minor peak in our function, whilst the main peak is reasonable for a tropical tree. However, the reason for the minimum of vegetation carbon in mid-latitudes is a prevalence of grasslands in drier conditions, which is partly captured by our moisture dependence of photosynthesis.

We define a vegetation fraction, $f_v$, as a saturating function of vegetation carbon that varies between 0 and 1. $f_v \to 1$ implies a closed canopy. A value of 0.5 implies half the areal view is vegetated.

$$f_v = 1 - e^{-k_{17}C_v} \quad (21)$$

We obtain a value for $k_{17}$ (Table 1) from a regression of vegetation fraction and vegetation carbon datasets (Figure A.4). Our approach is equivalent to a dependence of canopy light extinction on leaf area index, which in turn depends on biomass, as in e.g. Cox (1998). Both approaches make photosynthesis a saturating function of biomass.

Vegetation respiration depends on the air temperature and the amount of biomass, following Lenton (2000).

$$R_v = \frac{k_{24}}{k_{25}} f_4(T_a)C_v \quad (22)$$

Where $k_{24}$ is a vegetation respiration rate and $k_{25} = e^{\frac{-k_{20}}{R T_{\text{ref}}}}$ is a normalizing constant.

$$f_4(T_a) = e^{\frac{-k_{20}}{R T_a}} \quad (23)$$

Where $k_{20}$ is an activation energy, and $R = 8.314 \text{ J mol}^{-1} \text{ K}^{-1}$ is the universal gas constant.

The quantity of carbon lost by the vegetation to the soil via leaf litter, $L$, is related to the amount of vegetation carbon and the net primary productivity
(NPP), \( P - R_v \).

\[
L = k_{26}C_v + \varepsilon(P - R_v)
\]  

(24)

Where \( k_{26} \) is a turnover rate and \( \varepsilon \) represents self shading, given by

\[
\varepsilon = \frac{1}{1 + e^{k_{16} - C_v}}
\]  

(25)

This equation, inspired by TRIFFID (Cox, 1998), implies that all new production is channeled into leaf litter when \( \varepsilon \rightarrow 1 \). This occurs when \( f_v \rightarrow 1 \) i.e. the canopy is closed.

Carbon lost to the atmosphere by soil respiration, \( R_s \) is dependent on the land temperature \( T_l \) and the amount of carbon in the soil reservoir, \( C_s \).

\[
R_s = \frac{k_{29}}{k_{30}} f_1(T_l) C_s
\]  

(26)

Where \( k_{29} \) is the soil respiration rate and \( k_{30} \) is a normalizing constant. Above freezing (\( i = 5 \)), for \( T_l \geq 273.15 \) K

\[
f_5(T_l) = e^{-k_{31}/(T_l-k_{32})}
\]  

(27)

from Lloyd and Taylor (1994). Below freezing (\( i = 6 \)) the soil respiration rate is assumed to have a constant \( Q_{10} \) temperature sensitivity, for \( T_l < 273.15 \) K:

\[
f_6(T_l) = k_0 Q_{10}^{0.1(T_l-T_0)}
\]  

(28)

Where \( T_0 = 273.15 \) K, \( k_0 = f_5(T_0) \), and \( Q_{10} = e^{10k_{31}/(T_0-k_{32})^2} \). This prevents unrealistic blow-up of the soil reservoir as \( T_l \) approaches \( k_{32} \) in \( f_5 \).

2.4 Land albedo

The terrestrial surface albedo is dependent on what the surface is i.e. snow, vegetation, bare soil or sand, and is a function of vegetation and soil carbon. For a snow free surface, the terrestrial surface albedo is

\[
\alpha_s = f_v \alpha_v + (1 - f_v) \alpha_{soil}
\]  

(29)

where \( \alpha_v = 0.1 \) is the albedo of vegetation. The dependence of surface albedo on vegetation fraction \( (f_v) \) is similar to that in TRIFFID (Cox, 1998), where surface albedo is a saturating function of leaf area index, which in turn depends on biomass. \( \alpha_{soil} \) is given by

\[
\alpha_{soil} = \max\{\alpha_{peat}, (\alpha_{peat} - \alpha_{sand})\frac{k_{10}C_s}{k_8 - k_9} + \alpha_{sand}\}
\]  

(30)
Where $\alpha_{\text{peat}} = 0.11$ and $\alpha_{\text{sand}} = 0.3$. If snow is present the terrestrial surface albedo is calculated as

$$\alpha_s^{\text{snow}} = (\alpha_v^{\text{snow}} - \alpha_v) e^{-k_r C_v} + \alpha_v$$

(31)

Where $\alpha_v^{\text{snow}} = 0.3$ is the snow covered vegetation albedo and $\alpha_v^{\text{snow}} = 0.8$ is the albedo of a snow covered flat surface. We take the albedo values from Essery et al. (2001).

We assume any precipitation falling on a grid box falls as snow when both $T_a$ and $T_l$ are $\leq -5^\circ C$, as in Weaver et al. (2001). Our snow has zero thickness and therefore albedo due to snow is independent of snow depth, whereas other more detailed descriptions take snow depth into account (Essery et al., 2001; Dickinson et al., 1986; Petoukhov et al., 2000). Snow remains on a grid box until either $T_a$ or $T_l$ exceeds $-5^\circ C$. We attempt to parameterise only the leading order albedo effect of snow cover and hence neglect a full thermodynamic calculation that would include freezing, melting and sublimation.

### 3 Model tuning

The simplicity of the model allows very quick and easy tuning of the land carbon cycle when forced with fields of prescribed atmospheric data. These were long term monthly mean NCEP reanalysis air temperature, precipitation and relative humidity re-gridded data. We tuned parameters of the land carbon cycle model while holding the atmospheric concentration of $\text{CO}_2$ at a pre-industrial value of 278 ppmv. In this case the carbon equations become approximately linear and can be tuned to any value using only a few (2–4) runs of the model. Here we choose to tune the global annual average carbon fluxes of net photosynthesis, vegetation respiration, leaf litter and soil respiration to IPCC (Houghton et al., 2001) pre-industrial figures of 120 GtC yr$^{-1}$ (giga tonnes of carbon per year) for net photosynthesis and 60 GtC yr$^{-1}$ for the other fluxes, by altering the rate constants $k_{18}$, $k_{24}$, $k_{26}$ and $k_{29}$. These are then fine tuned in an iterative trial and error method to correct for the small non-linearities, giving the values in Table 1. We also tune global annual average values of vegetation and soil carbon toward the values of 451 GtC for vegetation and 1306 GtC for soil from Olson et al. (1985) and Batjes (1995) respectively. These data include effects of land use change, which have tended to reduce carbon storage, whereas ENTS calculates potential vegetation cover in the absence of land use change.

Our model gives global figures of 119.2 GtC yr$^{-1}$ for photosynthesis, 57.9 GtC yr$^{-1}$ for vegetation respiration, and 61.3 GtC yr$^{-1}$ for leaf litter and soil respiration. Global vegetation and soil carbon are 437 GtC and 1317 GtC
respectively. The response of the model equations to changing climate is tested by the quality of fit to spatial observations, because each spatial grid point has a different combination of climate drivers. However, the response to changing CO$_2$ and the parameters controlling it are not tuned here.

4 Results and comparison with data

Here we have forced the model with NCEP fields of air temperature, wind speed and humidity hence all the physical and biological land parameters are interactive whilst the atmospheric ones are prescribed.

4.1 Land temperature and hydrology

ENTS calculates land temperature as a function of the air temperature, the albedo and the evapotranspiration. The model predicted land temperature compares well with the NCEP field of skin temperature in most locations (Figure A.5 a-b). There are some large differences on some grid points, notably in Antarctica and Greenland. These are due to mismatches in the model and NCEP land-sea masks causing the regridding of the NCEP data to contain a fraction of sea properties. The generally excellent agreement supports the validity of our approximation to the solution of the land temperature (see Appendix).

The evaporation field produced by the model also compares well to NCEP reanalysis (Figure A.5 c-d). This suggests that our roughness length parameterisation is reasonable. However, there is a slight tendency to overestimate evaporation. As with the regridding of the NCEP skin temperature we find anomalous points in Antarctica, Greenland and the Western Pacific islands that can be explained from the mismatch of the land-sea masks.

We have not been able to find a global data set with which to compare the model results for fractional soil saturation, $W_s/W_{s^*}$ (Figure A.6). The Sahara stands out as the driest region, with other deserts having some soil moisture in the annual average. Tropical rainforest soils are close to field capacity whilst temperate and boreal forest soils tend to be somewhat below field capacity.

4.2 Vegetation and soil carbon

Comparing with vegetation carbon storage data (Figure A.7 a-b), the peak values in tropical rainforests and a secondary maximum in boreal forests are
captured, as are the main desert regions of the world. Peaks in boreal vegetation carbon are rather low and/or misplaced, but are much closer to the data than they would be without the secondary peak in photosynthesis at low temperature (Figure A.3). The effects of not accounting for land use change can be seen in China and Western Europe where the model predicts forests while the data shows lower figures for biomass. Model vegetation carbon is also too high in SE Brazil, whereas in Australia it is too low.

Comparing with soil carbon storage data (Figure A.7 c-d), the boreal peak values of soil carbon storage and modest values through the tropics are captured. The broad agreement is largely due to having a temperature sensitivity of soil respiration that is greater at lower temperatures, rather than using a fixed $Q_{10}$ response everywhere, as in e.g. TRIFFID (Cox, 1998). Boreal peaks in soil carbon are misplaced partly because we don’t include moisture control on soil respiration, which gives water-logged peatlands very high soil carbon. Neither do we explicitly represent permafrost, although we capture the very slow rates of decomposition at sub-zero temperatures. There is an erroneous peak in soil carbon in SE Asia, partly linked to not accounting for land use change. Amazon soil carbon is also a little too high. Soil carbon is too low in Australia, Southern Africa, the southern tip of South America, and other mid-latitude grassland regions. This is a weakness of the model not distinguishing the grass functional type, which despite its low biomass can support relatively high soil carbon.

Roughness length (not shown) is a linear function of vegetation carbon hence it is a scaled version of the model results in figure A.7a with a maximum of circa 0.5 m. This maximum is rather low for mature forests, especially ones with clearings, which is a limitation of our simple parameterisation (Figure A.2).

4.3 Vegetation fraction and land albedo

Our parameterisation of vegetation fraction compares well with observations, which show high coverage outside desert regions (Figure A.8 a-b). There is a tendency toward no vegetation in some regions which actually have reasonable fractional coverage, e.g. Australia. Owing to the clustering of values around 1 or 0 in the data and the model, a more rigorous look at the parameterisation is made difficult.

Vegetation fraction exerts a strong inverse control on land surface albedo in the absence of snow (Figure A.8 c-d). The observations of Matthews (1985) do not include the effects of snow on surface albedo whereas the model does. Hence we have removed the effects of snow and show the model land albedo
as a function of only the soil and vegetation albedos. Land surface albedo is accurate for deserts, but generally a little too low in other regions. This is partly because we only include one generic vegetation type with a minimum albedo of 0.1, appropriate for needleleaf and broadleaf trees but not grasses (Essery et al., 2001). When seasonal snow cover is included, this gives much higher land surface albedo values in boreal regions (not shown).

4.4 Net primary production

Net primary production (NPP), \( P - R_{v} \) in ENTS is compared with the average of the 17 models in the intercomparison of Cramer et al. (1999) in Figure A.9. Global NPP in ENTS is 61.3 GtC yr\(^{-1}\), which is towards the upper end of the range of 44.3–66.3 GtC yr\(^{-1}\) in the intercomparison (after two outliers had been removed) (Cramer et al., 1999), but is lower than the 70 GtC yr\(^{-1}\) in VECODE (Brovkin et al., 1997). Spatial peak values of NPP in ENTS are generally higher than the 17 model average, but similar to those in IBIS (Foley et al., 1996). Peak NPP is high relative to other models partly because we tune to a global total NPP of 60 GtC yr\(^{-1}\) from the IPCC, whereas satellite derived estimates are nearer to 50 GtC yr\(^{-1}\) (Field et al., 1998). Furthermore, our generic vegetation type does not capture aridity tolerant grasses, and combined with the coarseness of our grid this leaves large areas unvegetated. Hence in the remaining areas, NPP has to be higher to reach the global total NPP tuning target. More conservative figures for NPP could easily be obtained by re-tuning the model to different targets following the method described in Section 3.

5 Conclusion

We have described and tuned a new land surface physics and terrestrial carbon cycle model and tested it against data. ENTS is a minimal spatial model in terms of complexity and computational time, the logic being that this enables longer periods of earth history and larger ensembles to be simulated whilst also making diagnosis of the model processes easier. We started with the land carbon equations of simple zero dimensional box models described in Lenton (2000) and Adams (2003) and refined and modified them for use in a spatial model, developing new parameterisations for roughness length and vegetation fraction as simple functions of vegetation carbon density.

We created the model with a single soil layer and have shown that with the use of a diurnally averaged solar insolation forcing function, solving for the equilibrium solution of the land radiation balance is a valid approximation.
Using this approximation makes our model computationally undemanding as we can calculate the land temperature at a lower frequency than the usual finite difference time stepping approach. However, this approximation is not suitable for models which do resolve diurnal solar insolation.

ENTS has been incorporated into an Earth system model of intermediate complexity (EMIC), GENIE-1, with a closed carbon cycle (Lenton et al., 2006). This includes a 2-D energy moisture balance atmosphere, sea-ice, a 3-D ocean circulation model, GOLDSTEIN (Edwards and Marsh, 2005), and an ocean biogeochemistry model, BIOGEM (Ridgwell, 2001), and is being used for paleoclimate simulations and long-term future projections (Lenton et al., 2006). Ideally the model pre-industrial state should be re-tuned toward larger values for global vegetation and soil carbon storage. For simulations of present and future time a parameterisation of land use change still needs to be developed. The consequences of using a single soil carbon pool also need to be considered, as it cannot simultaneously capture the dynamics of short-lived detritus, mobile soil, and long-lived resistant soil (Lenton, 2000).

The fully coupled model can achieve up to 5000 years of integration per CPU hour, which puts $10^6$ yr model runs within reach. This raises the question of adaptation of vegetation. In common with all DGVMs and most ecosystem models, our model parameters are assumed constant throughout simulations. However, real vegetation exhibits considerable phenotypic plasticity and can also evolve on timescales of less than a million years. In model terms this means many of the parameters may vary on long timescales. This poses a conceptual challenge for future work.

6 Acknowledgements

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A Solving for the land temperature

In GENIE-1, all fluxes and the radiative forcing are calculated every 3.65 days of integration (the time step of the GOLDSTEIN ocean model). Hence
the radiation scheme does not resolve the diurnal cycle. If the land temperature approaches equilibrium in 3.65 days then we can solve directly for the equilibrium land temperature and thus increase the computational efficiency of ENTS. Here we consider how long it takes for the land to reach an equilibrium temperature by obtaining an analytic solution to the model’s radiation balance without the diurnal cycle. This prompted us to also find an analytic solution with a diurnal forcing term.

A.1 Land radiation balance

For our single soil layer, the time evolution of land temperature is given by:

\[ h_l c_l \frac{dT_l}{dt} = (1 - \alpha_{atm})(1 - \alpha_l)(1 - C_A)Q_{SW} - Q_{LH} - Q_{LW} - Q_{SH} \]  

(A.1)

On the left hand side of equation (A.1), \( h_l \) is a depth scale for land which is small, \(< 5 \) m and \( c_l = 3.3 \times 10^{-5} \) J m\(^{-3}\) K\(^{-1}\) is the heat capacity for land. For definition of the terms see section 2.1.

The radiative forcing term \( Q_{SW} \) depends on the average flux of radiation at the top of the atmosphere i.e. \( \overline{Q_{SW}} = S_\odot/4 \) where \( S_\odot = 1368 \) W m\(^{-2}\) is the solar constant. If one does not include a diurnal cycle (as in GENIE) and neglects the seasonal cycle (included in GENIE) and longer period astronomical terms such as Milankovitch cycles, then \( Q_{SW} \) is a constant for a fixed point on the Earth’s surface (section A.2). A simple diurnal cycle (section A.3) can be introduced using

\[ Q_{SW}(t) = \max\{0, \Re(Q_0 e^{i\Omega t})\} \]  

(A.2)

Where \( Q_0 \) is the amplitude of the incoming shortwave radiation, \( \Omega \) is the angular frequency of the Earth and \( \Re(z) \) denotes the real part of \( z \). This function would correspond to the average diurnal cycle at the equator as the length of the day is equal to the length of the night. To obtain different times for sunrise and sunset, equation (A.2) could be modified by including an additional \(+ c\) term, where \( c \) is a constant.

Figure A.1 shows the land-atmosphere radiation balance considered in this model where \( Q_{PLW} \) is the outgoing longwave planetary radiation and \( Q_T \) represents advection and diffusion of heat. In the analysis that follows, \( T_a \) is assumed constant. We can make this approximation provided \( T_a \) changes slowly relative to the land temperature \( T_l \), an assumption which is verified below. In GENIE-1 this approximation is valid because \( T_a \) is constant while the land radiation balance is calculated.
A.2 Solution without a diurnal cycle

We now obtain an analytical solution to equation (A.1). To make analytic solution possible we linearize $Q_{LW}$. To further simplify it, we also make the approximation that $\varepsilon_l \approx \varepsilon_a$. This gives $Q_{LW}$ as

$$Q_{LW} = B'(T_l - T_a)$$  \hspace{1cm} (A.3)

Where $B' = 5.13 \text{ W m}^{-2} \text{ K}^{-1}$ is the gradient of the linearized function.

We also initially make the approximation that there is no evaporation and hence no latent heat flux. This is true in desert regions or in situations when $q_a \geq q_v(T_l)$. Equation (A.1) now reduces to

$$h_l c_l \frac{dT_l}{dt} = Q' - \frac{1}{\eta} (T_l - T_a)$$  \hspace{1cm} (A.4)

Where $Q' = (1 - \alpha_{atm})(1 - \alpha_l)(1 - C_A)Q_{SW}$ is a constant with time, $t$, and temperature. $\eta = 1/(\rho_a C_H c_p a U + B')$ is a reciprocal measure of how effective the heat flux terms $Q_{SH}$ and $Q_{LW}$ are at transferring heat across the land-air interface.

The solution to equation (A.4) is

$$T_l(t) = T_0 e^{\frac{-t}{h_l c_l \eta}} + (1 - e^{\frac{-t}{h_l c_l \eta}})(T_a + \eta Q')$$  \hspace{1cm} (A.5)

Showing how the land temperature, $T_l$, evolves from an initial temperature $T_0$. From this solution we can now define an adjustment ($e$-folding) time scale, $t_{adj}$, as the time it takes for the initial temperature, $T_0$, to reach $1/e$ of its final value, $T_{eqm}$. From equation (A.5)

$$t_{adj} = h_l c_l \eta$$  \hspace{1cm} (A.6)

The equilibrium solution occurs when $Q' = Q_{LW} + Q_{SH}$. Taking the limit $t \to \infty$, $T_l \to T_{eqm}$ of equation (A.5) we have

$$T_{eqm} = T_a + \eta Q'$$  \hspace{1cm} (A.7)

As an illustrative case (Figure A.10a), for $T_0 = 20^\circ C$, $T_a = 10^\circ C$, $h_l = 5 \text{ m}$ and $U = 3 \text{ m s}^{-1}$, $t_{adj} \approx 0.54 \text{ days}$ and $T_{eqm} \approx 14^\circ C$.

Thus far we have assumed no latent heat flux, $Q_{LH}$. To obtain a solution with a latent heat flux we linearize $E$. This gives a timescale of:

$$t_{adj} = \frac{h_l c_l}{\rho_a C_H c_p a U + B' + C'(L \nu \rho_a C_W U)}$$  \hspace{1cm} (A.8)
Where $C' = 7.5 \times 10^{-4} \, ^\circ C^{-1}$ is a linearizing constant and $L_v = 2.5 \times 10^6 \, J \, \text{kg}^{-1}$ is the latent heat of vaporization. When evaporation is present, $t_{adj}$ is reduced, i.e. equilibrium is reached faster. For the same values as before, $t_{adj} \approx 0.21$ days.

### A.3 Solution with a diurnal cycle

Now we solve equation (A.4) but with $Q_{SW}$ given by equation (A.2) instead of a constant. We make the same approximation for $Q_{LW}$ (equation (A.3)) and again assume no evaporation so $Q_{LH} = 0$. This gives

$$h_l c_l \frac{dT_l}{dt} = Q' e^{i\Omega t} - \frac{1}{\eta}(T_l - T_a) \tag{A.9}$$

Where $Q' = (1 - \alpha_l)(1 - C_A)Q_0$ in this case. The solution is given by

$$T_l(t) = T_0 e^{\frac{t}{\tau_0}} + (1 - e^{\frac{t}{\tau_0}})[T_a + \eta Q' K(t)] \tag{A.10}$$

Where $K(t)$ is

$$K(t) = \max \left\{ 0, \frac{\cos(\Omega t) + \Omega h_l c_l \eta \sin(\Omega t)}{1 + \Omega^2 h_l^2 c_l^2 \eta^2} \right\} \tag{A.11}$$

It is worth comparing this solution, equation (A.10) with the constant forcing solution, equation (A.5). The time scale of adjustment, $t_{adj}$ is still given by equation (A.6) although this is the adjustment from the initial perturbation $T_0$. There is now a dynamic equilibrium temperature which can be seen by taking the limits $t \to \infty$, $T_l \to T_{eqm}$.

$$T_{eqm} = T_a + \eta Q' K(t) \tag{A.12}$$

In this dynamic equilibrium regime, we can define a phase difference, $\phi$, between the driving term, $Q_{SW} = \max\{0, \Re(Q_0 e^{i\Omega t})\} = \max\{0, Q_0 \cos(\Omega t)\}$ and the peak in the land temperature. If we define an angle, $\theta_1$, at which $Q_{SW}$ is maximized and an angle, $\theta_2$, at which $T_{eqm}$ is maximized then the phase difference, $\phi$, is

$$\phi = \theta_2 - \theta_1 \tag{A.13}$$

We define $\theta_1 = 0$ and this also has the absolute value $\theta_1 = \Omega t = 0$ as

$$\frac{dQ_{SW}}{dt} = Q_0 \Omega \sin(\Omega t) = 0 \tag{A.14}$$
which is also true for \( \Omega t = \theta = 0, 2\pi, \ldots \). The value of \( \theta_2 \) is given by \( \Omega t \) in the condition:

\[
\frac{dK(t)}{dt} = \Omega^2 h_l c_l \eta \cos(\Omega t) - \Omega \sin(\Omega t) = 0 \tag{A.15}
\]

This gives \( \theta_2 \) (ignoring the \( K = 0 \) night condition), as

\[
\theta_2 = \tan^{-1}(\Omega h_l c_l \eta) \tag{A.16}
\]

and \( \phi \) as

\[
\phi = \tan^{-1}(\Omega h_l c_l \eta) \tag{A.17}
\]

or in units of time, \( t_{lag} \) is

\[
t_{lag} = \frac{1}{\Omega} \tan^{-1}(\Omega h_l c_l \eta) \tag{A.18}
\]

We can look at the limits in this phase difference too. As \( \eta \to \infty \) which corresponds to slow transport of energy across the land-air interface then \( \phi \to \frac{\pi}{2} \) i.e. the land is 90° out of phase with the solar forcing. In other words the \( \sin(\Omega t) \) term dominates in the equation for \( K(t) \). In this case the difference between the maximum and minimum land temperatures over a diurnal cycle is very small as the denominator in \( K(t) \) (equation (A.11)) grows much quicker than the multiplier of the \( \sin(\Omega t) \) term. The other limit, very effective heat fluxes across the boundary, when \( \eta \to 0 \) then \( \phi \to 0 \) i.e. there is no phase difference between land response and driving term; the \( \cos(\Omega t) \) term in \( K(t) \) dominates. In this case the land can respond instantaneously to the forcing whereas in the former case the land experiences a lag due to the amount of heat it can hold. The magnitude of the change of \( T_l \) over a diurnal cycle is large as the denominator of \( K(t) \), equation (A.11), is small.

As an illustrative case (Figure A.10b), with the parameters used above (\( T_0 = 20^\circ C, T_a = 10^\circ C, h_l = 5 \text{ m} \) and \( U = 3 \text{ m s}^{-1} \)), \( t_{adj} \approx 0.54 \) days, as before, and \( \phi = 1.26 \) rad or 72 degrees.

### A.4 Sensitivity analysis

These parameter choices tend to overestimate the value of \( t_{adj} \). The depth scale of the land, \( h_l \), is more likely to be in the region of < 1 m and the magnitude of the wind speed, \( U \), has an annual average value of 4.9 m s\(^{-1}\) over land according to NCEP reanalysis data. We plot \( t_{adj} \) as a function of \( U \) in Figure A.11 with the parameters \( T_0 = 20^\circ C, T_a = 10^\circ C, h_l = 5 \text{ m} \). As a comparison, we also plot \( t_{adj} \) with evaporation included (equation (A.8)). From this it is clear that the land generally reaches an equilibrium temperature within 3 days. Hence we are able to solve for the equilibrium solution for \( T_l \), \( \frac{dT_l}{dt} = 0 \) in
equation (A.1), numerically via a fast Newton-Raphson root finding algorithm. However, if we included diurnal forcing in GENIE-1 then this method would not be applicable.

References


present climate. Climate Dynamics 16, 1–17.


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Table A.1
Parameter settings used in the model runs. $k_{18}$, $k_{24}$, $k_{26}$, and $k_{29}$ are tuned as described in Section 3.
Fig. A.1. A schematic of the land-atmosphere radiation balance, used in the Appendix to consider the equilibration of land temperature.
Fig. A.2. Vegetation carbon (kgC m$^{-2}$) from Olson et al. (1985) plotted against roughness length derived from NCEP reanalysis (points). Also plotted is the best linear fit to the data which is our parameterisation (line).
Fig. A.3. The temperature response of photosynthesis, $f_3$ (solid line) in ENTS, combining two functions, $f_{3a}$ (dotted line) and $f_{3b}$ (dashed line) in order to capture the responses of both low and high latitude vegetation types.
Fig. A.4. Vegetation carbon (kgC m$^{-2}$) from Olson et al. (1985) plotted against vegetation fraction (points). Also plotted is our parameterisation (line).
Fig. A.5. Annual mean land temperature (°C) in (a) ENTS model (b) NCEP data. Annual mean evaporation (m s⁻¹) in (c) ENTS model (d) NCEP data.
Fig. A.6. Annual average fractional soil saturation from the ENTS model.
Fig. A.7. Annual average vegetation carbon (kg C m$^{-2}$) in (a) ENTS model forced off-line with NCEP reanalysis data (b) Observations (Olson et al., 1985). Annual average soil carbon (kg C m$^{-2}$) in (c) ENTS model (d) Observations (Batjes, 1995).
Fig. A.8. Annual mean vegetation fraction (a) ENTS model (b) Observations (Hall et al., 2005). Annual average surface albedo (c) ENTS model forced with NCEP reanalysis data, excluding the effects of snow (d) Observations (Matthews, 1985), excluding the effects of snow.
Fig. A.9. Annual average net primary production (NPP) (a) from the ENTS model and (b) the average of 17 NPP models from the intercomparison of Cramer et al. (1999). Units are kg C m$^{-2}$ yr$^{-1}$. 

\[ \frac{\text{(b)}}{32} \]
Fig. A.10. Land temperature evolution, $T_L$ (a) without a diurnal cycle, (b) with a diurnal cycle. Parameter settings are: $T_0 = 20^\circ$C, $T_a = 10^\circ$C, $h_l = 5$ m and $U = 3$ m s$^{-1}$. This gives $t_{adj} \approx 0.54$ days in both cases. In (a) $T_{eqm} \approx 14^\circ$C, and in (b) $\phi = 1.26$ rad or 72 degrees.
Fig. A.11. The adjustment time scale for land, $t_{adj}$ as a function of the magnitude of the wind speed, $U$. Responses with and without evaporation ($Q_{LH}$) are shown. Parameters for this plot are: $T_0 = 20^\circ C$, $T_a = 10^\circ C$ and $h_l = 5$ m.