

Conditions for Positive Feedbacks from the Land Carbon Cycle

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Abstract

The first GCM climate-carbon cycle simulations have indicated that the land biosphere could provide a significant positive feedback on 21st century climate change (Cox *et al* (2000)). The positive feedback occurs in these numerical experiments because land carbon storage is projected to decrease from about 2050 onwards as temperature-enhanced respiration overwhelms CO₂-enhanced photosynthesis. The critical atmospheric CO₂ concentration at which the land in each GCM gridbox switches from carbon sink to carbon source can be understood in terms of a simple land carbon balance model, with effective parameters representing the sensitivities of climate and photosynthesis to CO₂, and the sensitivities of soil respiration and photosynthesis to temperature. This model is used to show that a carbon sink-to-source transition is inevitable beyond some finite CO₂ concentration provided a few simple conditions are satisfied.

1 Introduction

The first General Circulation Model (GCM) experiments to include the carbon cycle as an interactive element suggest that feedbacks between the climate and the land biosphere could significantly accelerate atmospheric CO₂ rise and climate change over the 21st century (Cox *et al* (2000)). These numerical experiments demonstrate the potential importance of climate-carbon cycle feedbacks, but the magnitude of these in the real Earth system is still highly uncertain. The strongest feedbacks, and therefore the greatest uncertainties, seem to be associated with the terrestrial biosphere. The cause of the present day land carbon sink is still in doubt, with CO₂-fertilisation, nitrogen deposition and forest regrowth all implicated in certain regions. The location of this sink is even more debatable, perhaps because this is subject to great interannual variability. Whilst increases in atmospheric CO₂ are expected to enhance photosynthesis (and reduce transpiration), the associated climate warming is likely to increase plant and soil respiration. Thus there is a competition between the direct effect of CO₂, which tends to increase terrestrial carbon storage, and the indirect effect, which may reduce carbon storage.

The outcome of this competition has been seen in a range of DGVMs (Cramer *et al* (2001)), each of which simulate reduced land carbon under climate change alone and increased carbon storage with CO₂ increases only. In most DGVMs, the combined effect of the CO₂ and associated climate change results in a reducing sink towards the end of the 21st century, as CO₂-induced fertilisation begins to saturate but soil respiration continues to increase with temperature. The manner in which soil and plant respiration respond in the long-term to temperature is a key uncertainty in the projections of CO₂ in the 21st century (Giardina and Ryan (2000)).

2 Conditions for Positive Feedback from the Land Carbon Cycle

In this sub-section we introduce a simple terrestrial carbon balance model to demonstrate how the conversion of a land CO₂ sink to a source is dependent on the responses of photosynthesis and respiration to CO₂ increases and climate warming. We consider the total carbon stored in vegetation and soil, C_T , which is increased by photosynthesis, Π , and reduced by the total ecosystem respiration, R :

$$\frac{dC_T}{dt} = \Pi - R \quad (1)$$

where Π is sometimes called Gross Primary Productivity (GPP), and R represents the sum of the respiration fluxes from the vegetation and the soil. In common with many others (McGuire *et al* (1992), Collatz *et al* (1991), Collatz *et al* (1992), Sellers *et al* (1996), Cox *et al* (1998)), we assume that GPP depends directly on the atmospheric CO₂ concentration, C_a , and the surface temperature, T (in °C):

$$\Pi = \Pi_{max} \left\{ \frac{C_a}{C_a + C_{0.5}} \right\} f(T) \quad (2)$$

where Π_{max} is the value which GPP asymptotes towards as $C_a \rightarrow \infty$, $C_{0.5}$ is the ‘‘half-saturation’’ constant (i.e. the value of C_a for which Π is half this maximum value), and $f(T)$ is an arbitrary function of temperature. We also assume that the total ecosystem respiration, R , is proportional to the total terrestrial carbon, C_T . The specific respiration rate (i.e. the respiration per unit carbon) follows a ‘‘Q10’’ dependence, which means that it increases by a factor of q_{10} for a warming of T by 10°C. Thus the ecosystem respiration rate is given by:

$$R = r C_T q_{10}^{(T-10)/10} \quad (3)$$

where r is the specific respiration rate at $T = 10^\circ\text{C}$. It is more usual to assume separate values of r and q_{10} for different carbon pools (e.g. soil/vegetation, leaf/root/wood), but our simpler assumption will still offer good guidance as long as the relative sizes of these pools do not alter significantly under climate change. Near surface temperatures are expected to increase approximately logarithmically with the atmospheric CO₂ concentration, C_a (Huntingford and Cox (2000)):

$$\Delta T = \frac{\Delta T_{2 \times CO_2}}{\log 2} \log \left\{ \frac{C_a}{C_a(0)} \right\} \quad (4)$$

where ΔT is the surface warming, $\Delta T_{2 \times CO_2}$ is the climate sensitivity to doubling atmospheric CO₂, and $C_a(0)$ is the initial CO₂ concentration. We can use this to eliminate CO₂ induced temperature changes from equation 3:

$$R = r_0 C_T \left\{ \frac{C_a}{C_a(0)} \right\}^\alpha \quad (5)$$

where $r_0 C_T$ is the initial ecosystem respiration (i.e. at $C_a = C_a(0)$) and the exponent α is given by:

$$\alpha = \frac{\Delta T_{2 \times CO_2}}{10} \frac{\log q_{10}}{\log 2} \quad (6)$$

We can now use equations 1, 2 and 5 to solve for the equilibrium value of terrestrial carbon, C_T^{eq} :

$$C_T^{eq} = \Pi_{max} \left\{ \frac{C_a}{C_a + C_{0.5}} \right\} \left\{ \frac{C_a(0)}{C_a} \right\}^\alpha \frac{f(T)}{r_0} \quad (7)$$

The land will tend to amplify CO₂-induced climate change if C_T^{eq} decreases with increasing atmospheric CO₂ (i.e. $dC_T^{eq}/dC_a < 0$). Differentiating equation 7 with respect to C_a yields:

$$\frac{dC_T^{eq}}{dC_a} = C_T^{eq} \left[\frac{(1 - \alpha_*)}{C_a} - \frac{1}{C_a + C_{0.5}} \right] \quad (8)$$

where:

$$\alpha_* = \frac{\Delta T_{2 \times CO_2}}{\log 2} \left\{ \frac{\log q_{10}}{10} - \frac{1}{f} \frac{df}{dT} \right\}. \quad (9)$$

The condition for the land to become a source of carbon under increasing CO₂ is therefore:

$$C_a > \frac{1 - \alpha_*}{\alpha_*} C_{0.5} \quad (10)$$

This means that there will always be a critical CO₂ concentration beyond which the land becomes a source, as long as:

- (i) CO₂ fertilisation of photosynthesis saturates at high CO₂, i.e. $C_{0.5}$ is finite.
- (ii) $\alpha_* > 0$, which requires:
 - (a) climate warms with increasing CO₂, i.e. $\Delta T_{2 \times CO_2} > 0$
 - (b) respiration increases more rapidly with temperature than GPP, i.e.

$$\frac{\log q_{10}}{10} > \frac{1}{f} \frac{df}{dT}. \quad (11)$$

Conditions (i) and (ii)(a) are satisfied in the vast majority of climate and terrestrial ecosystem models. Detailed models of leaf photosynthesis indicate that $C_{0.5}$ will vary with temperature from about 300 ppmv at low temperatures, up to about 700 ppmv at high temperatures (Collatz *et al* (1991)). Although there are differences in the magnitude and patterns of predicted climate change, all GCMs produce a warming when CO₂ concentration is doubled. The global mean climate sensitivity produced by these models ranges from 1.5K to 4.5K (Houghton *et al* (1996)), but mean warming over land is likely to be a more appropriate measure of the climate change experienced by the land biosphere. We estimate a larger range of $2K < \Delta T_{2 \times CO_2} < 7K$, because the land tends to warm more rapidly than the ocean (Huntingford and Cox (2000)).

There is considerable disagreement over the likely long-term sensitivity of respiration fluxes to temperature, with some suggesting that temperature-sensitive “labile” carbon pools will soon become exhausted once the ecosystem enters a negative carbon balance (Giardina and Ryan (2000)). However, condition (ii)(b) is satisfied by the vast majority of existing land carbon cycle models, and seems to be implied (at least on the 1-5 year timescale) by climate-driven interannual variability in the measured atmospheric CO₂ concentration (Jones and Cox (2001), Jones *et al* (2001)).

a Application to the Contemporary Climate

Most would therefore agree that the terrestrial carbon sink has a finite lifetime, but the length of this lifetime is highly uncertain. We can see why this is from our simple model (equation 10). The critical CO₂ concentration is very sensitive to α_* which is itself dependent on the climate sensitivity, and the difference between the temperature dependences of respiration and GPP (equation 9).

We expect the temperature sensitivity of GPP to vary regionally, since generally a warming is beneficial for photosynthesis in mid and high latitudes (i.e. $df/dT > 0$), but not in the tropics where the existing temperatures are near optimal for vegetation (i.e. $df/dT \leq 0$). As a result, we might expect global mean GPP to be only weakly dependent on temperature ($df/dT \approx 0$). We can therefore derive a range for α_* , based on plausible values of climate sensitivity over land ($2K < \Delta T_{2 \times CO_2} < 7K$) and respiration sensitivity ($1.5 < q_{10} < 2.5$). This range of $0.1 < \alpha_* < 0.9$, translates into a critical CO₂ concentration which is somewhere between 0.1 and 9 times the half-saturation constant (equation 10). Therefore on the basis of this simple analysis the range of possible critical CO₂ values spans almost 2 orders of magnitude. Evidently, the time at which the sink-to-source transition will occur is extremely sensitive to these uncertain parameters. This may explain why many of the existing terrestrial models do not reach this critical point before 2100 (Cramer *et al* (2001)).

Fortunately we can reduce the uncertainty range further. Critical CO₂ values which are lower than the current atmospheric concentration are not consistent with the observations, since the “natural” land ecosystems appear to be a net carbon sink rather than a source at this time (Schimel *et al* (1996)). For a typical half-saturation constant of $C_{0.5} = 500$ ppmv this implies that all combinations of q_{10} and $\Delta T_{2\times CO_2}$ which yield values of $\alpha_* < 0.6$ are unrealistic. Also, sensitivity tests with our coupled model indicate that $q_{10} = 2$ provides an almost optimal fit to the observed variability in atmospheric CO₂ due to ENSO (Jones *et al* (2001)) and volcanic eruptions (Jones and Cox (2001)), suggesting that the probability distribution for possible q_{10} values is peaked quite sharply about this value. Similarly the complete range of climate sensitivity values are not all equally probable, since the most advanced GCMs tend to produce values clustered around the centre of the range. It is therefore meaningful to produce a central estimate for the critical CO₂ value. Using $q_{10} = 2$, $C_{0.5} = 500$ ppmv, and $\Delta T_{2\times CO_2} = 4.8$ K (which is consistent with the warming over land in our coupled model,) yields a critical CO₂ value of about 550 ppmv, which is remarkably close to the sink-to-source transition seen in our experiment.

We draw two main conclusions from this section. The recognised uncertainties in climate and respiration sensitivity imply a very large range in the critical CO₂ concentration beyond which the land will act as a net carbon source. However, the central estimates for these parameters suggest a significant probability of this critical point being passed by 2100 in the real Earth system, under a “business as usual” emissions scenario, in agreement with the results from our coupled climate-carbon cycle model.

3 Conclusions

I have presented a simple analysis to demonstrate that a sink-to-source transition of the terrestrial biosphere is assured beyond some critical atmospheric CO₂ concentration, provided that a few simple conditions apply. Qualitatively then, the eventual saturation of the land carbon sink and its conversion to a carbon source, is supported by our existing understanding of terrestrial ecosystem processes.

Unfortunately, the precise point at which the land biosphere will start to provide a positive feedback cannot yet be predicted with certainty. This depends on a number of poorly understood processes, such as the long-term response of photosynthesis and soil respiration to increased temperatures (Giardina and Ryan (2000)), and the possible acclimation of photosynthesis to high CO₂. Our results suggest that accurate prediction of climate change over the 21st century, will be as dependent on advances in the understanding and modelling of these physiological and ecological processes, as it is on the modelling of the physical processes currently represented in GCMs.

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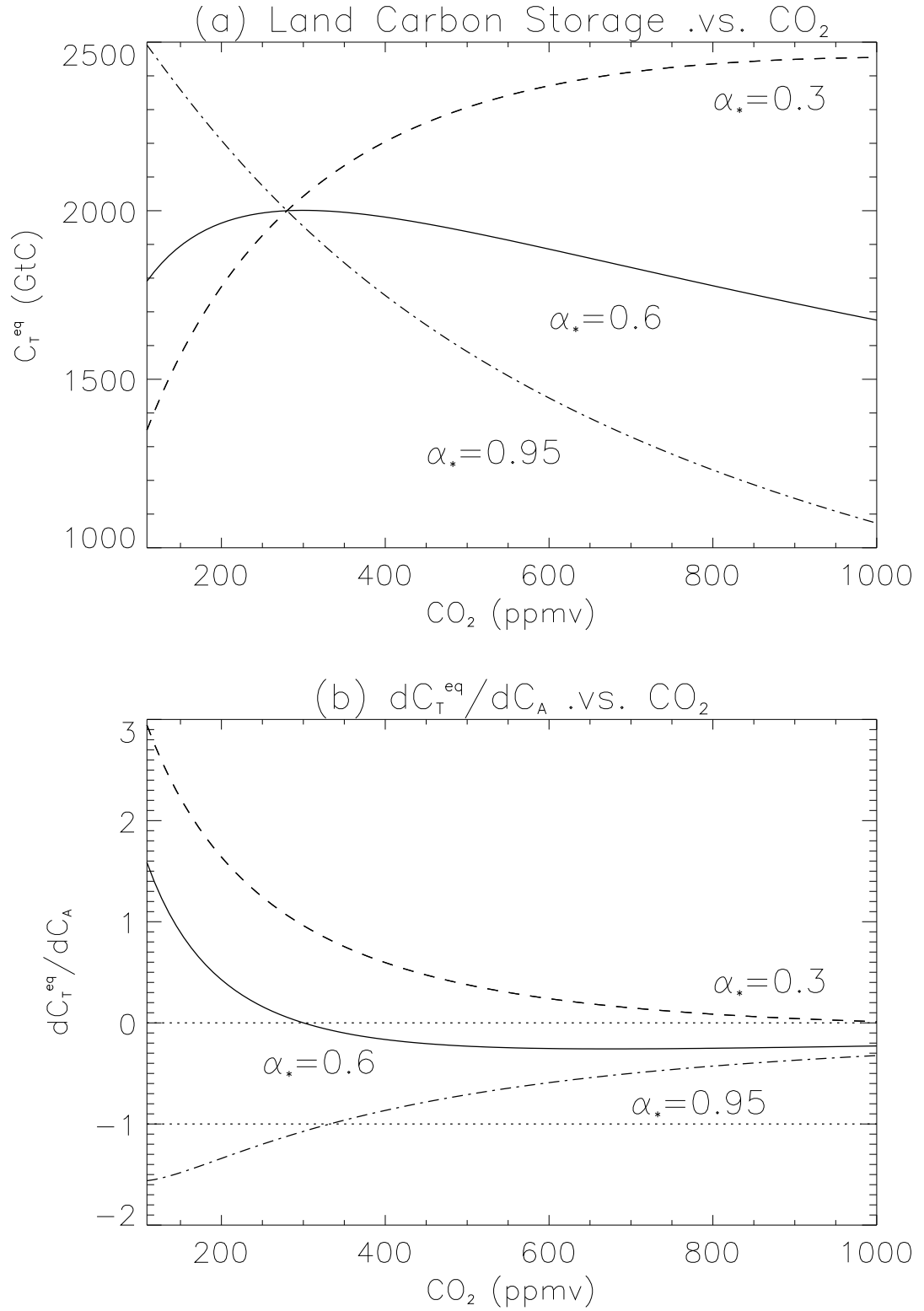


Figure 1: (a) Equilibrium Land carbon storage, C_T^{eq} , and (b) rate of change of equilibrium land carbon with respect to atmospheric carbon, dC_T^{eq}/dC_A , both versus atmospheric CO₂ concentration for three values of α_* . These curves are calculated from equations 7 and 8, assuming $C_a(0) = 280$ ppmv, $C_T(0) = 2000$ GtC, $\Pi(0) = 120$ GtC yr⁻¹, $C_{0.5} = 500$ ppmv, and $f(T) = 1$.