CHAPTER 15

Conditions for Sink-to-Source Transitions and Runaway Feedbacks from the Land Carbon Cycle

Peter M. Cox¹, Chris Huntingford² and Chris D. Jones³
¹Centre for Ecology and Hydrology, Winfrith, Dorset, UK
²Centre for Ecology and Hydrology, Wallingford, Oxon, UK
³Hadley Centre, Met Office, Fitzroy Road, Exeter, UK

ABSTRACT: The first GCM climate-carbon cycle simulation indicated that the land biosphere could provide a significant acceleration of 21st century climate change (Cox et al. 2000). In this numerical experiment the carbon storage was projected to decrease from about 2050 onwards as temperature-enhanced respiration overwhelmed CO₂-enhanced photosynthesis. Subsequent climate-carbon cycle simulations also suggest that climate change will suppress land-carbon uptake, but typically do not predict that the land will become an overall source during the next 100 years (Friedlingstein et al., accepted). Here we use a simple land carbon balance model to analyse the conditions required for a land sink-to-source transition, and address the question; could the land carbon cycle lead to a runaway climate feedback?

The simple land carbon balance model has 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) a carbon sink-to-source transition is inevitable beyond some finite critical CO₂ concentration provided a few simple conditions are satisfied, (b) the value of the critical CO₂ concentration is poorly known due to uncertainties in land carbon cycle parameters and especially in the climate sensitivity to CO₂, and (c) that a true runaway land carbon-climate feedback (or linear instability) in the future is unlikely given that the land masses are currently acting as a carbon sink.

15.1 Introduction

Vegetation and soil contain about three times as much carbon as the atmosphere, and they exchange very large opposing fluxes of carbon dioxide with it. Currently the land is absorbing about a quarter of anthropogenic CO₂ emissions, because uptake by plant photosynthesis is outstripping respiration from soils (Houghton et al. 1996). However, these opposing fluxes are known to be sensitive to climate, so the fraction of emissions taken up by the land is likely to change in the future. A number of authors have discussed the possibility of the land carbon sink either saturating or reversing (see for example Woodwell and Mackenzie (1995), Lenton and Huntingford (2003)), primarily because of the potential for accelerated decomposition of soil organic matter under global warming (Jenkinson et al. 1991). Simple box models of the climate carbon system have also demonstrated sink-to-source transitions in the land carbon cycle (e.g. Lenton 2000).

The General Circulation Models (GCMs) used to make climate projections have typically neglected such climate-carbon cycle feedbacks, but recently a number of GCM modelling groups have begun to include representations of vegetation and the carbon cycle within their models. The first GCM simulation of this type suggested 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). Subsequent GCM climate-carbon cycle projections also suggest that climate change will suppress land carbon uptake, but typically do not predict that the land will become a carbon source within the simulated period to 2100 (Friedlingstein, accepted).

The terrestrial components used in these first generation coupled climate-carbon cycle GCMs reproduce the land carbon sink as a competition between the direct effects of CO₂ on plant growth, and the effects of climate change on plant and soil respiration. 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 battle between the direct effect of CO₂, which tends to increase terrestrial carbon storage, and the indirect effect through climate warming, which may reduce carbon storage.

The outcome of this competition has been seen in a range of dynamic global vegetation models or ‘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. This is in itself an important result as it suggests that climate change will suppress the land carbon sink, and therefore lead to greater rates of CO₂ increase and global warming.
than previously assumed. However, in most models the land carbon cycle remains an overall sink for CO₂ and thus continues to provide a brake on increasing atmospheric CO₂.

The impact of climate change on the land carbon cycle is especially strong in the coupled model projections of Cox et al. (2000), leading to the land carbon cycle becoming an overall source of CO₂ from about 2050 onwards (under a ‘business as usual’ emissions scenario). In this case the land carbon cycle stops slowing climate change, and instead starts to accelerate it by releasing additional CO₂ to the atmosphere. This ‘sink-to-source’ transition point may be seen as one possible definition of ‘dangerous climate change’. In the next section of this chapter we use a transparently simple land carbon cycle model to derive a condition for the critical CO₂ concentration at which the sink-to-source transition will occur. The resulting analytical expression is used to highlight the key uncertainties that contribute to divergences amongst existing DGVM and GCM model projections (section 15.2.1).

Section 15.3 examines the conditions for an even stronger ‘runaway’ land carbon cycle feedback. In this case the carbon cycle-climate system becomes linearly unstable to an arbitrary perturbation, leading to a release of land carbon to the atmosphere even in the absence of anthropogenic emissions. This state therefore represents not just ‘dangerous climate change’ but ‘rapid climate change’ in which the CO₂ increase and climate change are potentially much faster than the rate of anthropogenic forcing of the system. We use the simple model to show that such a runaway feedback is possible in principle (e.g. if the climate sensitivity to CO₂ is very high), but is unlikely given the existence of a land carbon sink in the present day.

15.2 Conditions for Sink-to-Source Transitions in the Land Carbon Cycle

In this section we introduce a very 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, \( C_a \), which is increased by photosynthesis, \( \Pi \), and reduced by the total ecosystem respiration, \( R \):

\[
\frac{dC_a}{dt} = \Pi - R
\]  

(15.1)

where \( \Pi \) 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. 1998), we assume that GPP depends directly on the atmospheric CO₂ concentration, \( C_a \), and the surface temperature, T (in °C):

\[
\Pi = \Pi_{\text{max}} \left[ \frac{C_a}{C_a + C_{a,0.5}} \right] f(T)
\]  

(15.2)

where \( \Pi_{\text{max}} \) is the value which GPP asymptotes towards as \( C_a \to \infty \), \( C_{a,0.5} \) is the ‘half-saturation’ constant (i.e. the value of \( C_a \) for which \( \Pi \) 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:

\[
R = r C_T q_{10}^{(T_{90} - T_{10})/10}
\]  

(15.3)

where \( r \) is the specific respiration rate at \( T = 10°C \). 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, \( C_a \) (Houghton et al. 1996):

\[
\Delta T = \frac{\Delta T_{\text{CO₂}}}{\log 2} \log \left[ \frac{C_a}{C_a(0)} \right]
\]  

(15.4)

where \( \Delta T \) is the surface warming, \( \Delta T_{\text{CO₂}} \) 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 15.3:

\[
R = r_0 C_T \left[ \frac{C_a}{C_a(0)} \right]^\mu
\]  

(15.5)

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

\[
\mu = \frac{\Delta T_{\text{CO₂}}}{10} \log q_{10}^{(T_{90} - T_{10})/10} \log 2
\]  

(15.6)

We can now use Equations 15.1, 15.2 and 15.5 to solve for the equilibrium value of terrestrial carbon, \( C_T^{eq} \):

\[
C_T^{eq} = \Pi \left[ \frac{C_a}{C_a + C_{a,0.5}} \right] f(T_{90}) f(T_{10})
\]  

(15.7)

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 15.7 with respect to \( C_a \) yields:

\[
\frac{dC_T^{eq}}{dC_a} = C_T^{eq} \left[ (1 - \mu_*) - \frac{1}{C_a + C_{a,0.5}} \right]
\]  

(15.8)

where

\[
\mu_* = \frac{\Delta T_{\text{CO₂}}}{10} \log q_{10}^{(T_{90} - T_{10})/10} \frac{df}{dT}
\]  

(15.9)
The equilibrium land carbon storage, $C^{eq}$ (Equation 15.7), and the rate of change of equilibrium land carbon with respect to atmospheric carbon $dC^{eq}_L/dC_A$ (Equation 15.8), are plotted in Figures 15.1 and 15.2 for three values of $\mu_*$. For small values of $\mu_*$ the equilibrium land carbon increases monotonically over the range of CO$_2$ concentrations of interest (180–1000 ppmv), implying that the land would act as a carbon sink throughout the 21st century. By contrast, large values of $\mu_*$ show a monotonically decreasing land carbon storage with CO$_2$ concentration, implying a continuous land carbon source, which is at odds with the existence of a current-day land carbon sink. Only for intermediate values of $\mu_*$ do we see a turning point in the land carbon storage as a function of CO$_2$, with a current-day land carbon sink becoming a source before the end of the century (Figure 15.2).

![Figure 15.1](image1)

**Figure 15.1** Equilibrium land carbon storage, $C^{eq}_L$, versus atmospheric CO$_2$ concentration for three values of $\mu_*$. These curves are calculated from Equation 15.7 assuming $C_0(0) = 280$ ppmv, $C_T(0) = 2000$ GtC, $I(0) = 120$ GtC yr$^{-1}$, $C_{0.5} = 500$ ppmv, and $f(T) = 1$.

![Figure 15.2](image2)

**Figure 15.2** Rate of change of equilibrium land carbon with respect to atmospheric carbon, $dC^{eq}_L/dC_A$, versus atmospheric CO$_2$ concentration for three values of $\mu_*$. These curves are calculated from Equation 15.8 assuming $C_0(0) = 280$ ppmv, $C_T(0) = 2000$ GtC, $I(0) = 120$ GtC yr$^{-1}$, $C_{0.5} = 500$ ppmv, and $f(T) = 1$.

The sink-to-source turning point occurs where the rate of change of land carbon storage with CO$_2$ passes through zero, from positive (carbon sink), to negative (carbon source). From Equation 15.8, the condition for the land to become a source of carbon under increasing CO$_2$ is therefore:

$$C_a > \frac{1 - \mu_*}{\mu_*} C_{0.5}$$  \hspace{1cm} (15.10)

This means that there will always be a critical CO$_2$ concentration beyond which the land becomes a source, as long as:

(i) CO$_2$ fertilisation of photosynthesis saturates at high CO$_2$, i.e. $C_{0.5}$ is finite.

(ii) $\mu_* > 0$, which requires:
   (a) climate warms with increasing CO$_2$, i.e. $\Delta T_{2xCO_2} > 0$
   (b) respiration increases more rapidly with temperature than GPP, i.e.

$$\frac{\log q_{10}}{10} > \frac{df}{f dT}$$  \hspace{1cm} (15.11)

Conditions (i) and (ii)(a) are satisfied in the vast majority of terrestrial ecosystem and climate 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 CO$_2$ concentration is doubled.

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 inter-annual variability in the measured atmospheric CO$_2$ concentration (Jones and Cox 2001, Jones et al. 2001).

### 15.2.1 Application to the Contemporary Climate

We therefore conclude 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 15.10). The critical CO$_2$ concentration is very sensitive to $\mu_*$ which is itself dependent on the climate sensitivity, and the difference between the temperature dependences of respiration and GPP (Equation 15.9).

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. $df/dT \approx 0$). As a result,
we might expect global mean GPP to be only weakly dependent on temperature (d/dT \approx 0), even though there may be significant regional climate effects on GPP through changes in water availability.

Most climate models produce estimates of climate sensitivity to doubling CO₃ in the often-quoted range of 1.5 K to 4.5 K (Houghton et al. 1996), but there is now a growing realisation that the upper bound on climate sensitivity is much higher. A recent ‘parameter ensemble’ of GCM experiments (in which each ensemble member has a different set of feasible internal model parameters) produced model variants with climate sensitivities as high as 11 K (Stainforth et al. 2005). In principle it ought to be possible to estimate climate sensitivity by using the observed warming over the 20th century as a constraint. Unfortunately, in practice high climate sensitivities cannot be ruled out owing to uncertainties in the extent to which anthropogenic aerosols have offset greenhouse warming (Andreae et al. 2005).

In order to demonstrate the uncertainties in the critical CO₂ concentration we take the conservative 1.5 to 4.5 K range for the global climate sensitivity. 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 2 K \( \leq \Delta T_{2xCO₂} \leq 7 K \) because the land tends to warm more rapidly than the ocean (Huntingford and Cox 2000). The sensitivity of ecosystem respiration to temperature, as summarised by the \( q_{10} \) parameter, is known to vary markedly amongst ecosystems, but here we require an effective value to represent the climate sensitivity of global ecosystem respiration. Fortunately, anomalies in the growth-rate of atmospheric CO₂, associated with El Niño events (Jones et al. 2001), and the Pinatubo volcanic eruption (Jones and Cox 2001), give a reasonably tight constraint on this parameter of 1.5 \( < q_{10} < 2.5 \).

We can therefore derive a range for \( \mu_s \), based on plausible values of climate sensitivity over land (2 K \( \leq \Delta T_{2xCO₂} \leq 7 K \)) and respiration sensitivity (1.5 \( < q_{10} < 2.5 \)). This range of 0.1 \( < \mu_s < 0.9 \), translates into a critical CO₂ concentration which is somewhere between 0.1 and 9 times the half-saturation constant (Equation 15.10). Therefore on the basis of this simple analysis the range of possible critical CO₂ values spans almost two 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), Friedlingstein et al. 2005). It is also interesting to note that the ‘central estimate’ of \( q_{10} = 2 \), \( C_{A5} = 500 \) ppbv, and \( \Delta T_{2xCO₂} = 4.8 K \) (which is consistent with the warming over land in the Hadley Centre coupled model) yields a critical CO₂ value of about 550 ppbv, which is remarkably close to the sink-to-source transition seen in the Hadley Centre experiment.

In the absence of significant non-CO₂ effects on climate change (i.e. assuming that anthropogenic aerosols have approximately offset the warming due to the minor greenhouse gases), we can reduce the uncertainty range further. Under this assumption, 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 \( C_{A5} = 500 \) ppbv this implies that combinations of \( q_{10} \) and \( \Delta T_{2xCO₂} \) which yield values of \( \mu_s > 0.6 \) are unrealistic. We will return to this point in section 15.3.

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 real possibility of this critical point being passed by 2100 in the real Earth system, under a ‘business as usual’ emissions scenario, in qualitative agreement with the results from the Hadley Centre coupled climate-carbon cycle model.

15.3 Conditions for Runaway Feedback from the Land Carbon Cycle

The sensitivity of a system can be defined in terms of the relationship between the forcing of the system (e.g. anthropogenic CO₂ emissions) and its response (e.g. global warming). Rapid or abrupt change is normally associated with responses that are much larger than the forcing, or even independent of it. The latter are typically described as ‘instabilities’.

Although a sink-to-source transition in the land carbon cycle would imply an acceleration of climate change, it would not necessarily lead to a sudden change in the Earth System. In this section we examine the necessary conditions for the land carbon-climate system to be linearly unstable at some finite CO₂ concentration. If such a threshold existed, and was crossed, the land would spontaneously lose carbon to the atmosphere, leading to sufficient greenhouse warming to sustain the release even in the absence of anthropogenic emissions. Such instabilities are often termed ‘runaway feedbacks’ because of their self-sustaining nature.

Even such strong positive feedbacks are ultimately limited by the depletion of reservoirs (e.g. soil carbon), and longer-term negative feedbacks (e.g. uptake of CO₂ by the oceans). In the context of land carbon-climate feedbacks on the century timescale, fast carbon loss from the tropics may completely overwhelm slow carbon uptake in high latitudes, even though in the longer term the biosphere may contain more carbon under high CO₂ conditions. These very different timescales for carbon loss and accumulation mean that the existence of high-carbon storage on the land during hot climates of the past (e.g. the mid-Cretaceous 100 million years ago) does not rule out the possibility of transient runaway instabilities under anthropogenic climate change in the future.
A runaway condition is defined by an instability such that a small perturbation grows exponentially, i.e., a runaway positive feedback requires linear instability (i.e., a feedback gain factor greater than 1). Although Equation 15.10 defines the critical CO₂ concentration for the land carbon cycle to provide a positive feedback, it does not ensure that this feedback is strong enough for a runaway. In order to define the condition for linear instability we rewrite Equation 15.1 in the form:

\[
\frac{dC_T}{dt} = \frac{C_T^{eq} - C_T}{\tau} \tag{15.12}
\]

Here we have used Equation 15.5 to define the timescale, \(\tau\), which characterizes the rate at which the terrestrial carbon storage, \(C_T\), approaches its equilibrium value, \(C_T^{eq}\):

\[
\tau = \frac{1}{r_0} \left[ \frac{C_T^{eq}(0)}{C_a} \right]^{1/p} \tag{15.13}
\]

We consider a perturbation to an initial equilibrium state defined by \(C_T = C_T^{eq}(0)\) and \(C_a = C_a(0)\), where \(C_a\) is the atmospheric carbon content, in GtC, associated with the CO₂ concentration \(C_a\) in ppm (\(C_a = 2.123 C_a\)). A runaway occurs when \(C_a\) increases even in the absence of any CO₂ emissions, such that the total carbon in the atmosphere-land-ocean system is conserved:

\[
\Delta C_A + \Delta C_T + \Delta C_O = 0 \tag{15.14}
\]

where \(\Delta C_A\), \(\Delta C_T\) and \(\Delta C_O\) represent perturbations to the carbon in the atmosphere, land and ocean respectively. For simplicity we assume that the ocean takes up a fraction \(x_o\) of any increase in atmospheric carbon, i.e., \(\Delta C_o = x_o \Delta C_A\), so the carbon conservation Equation becomes:

\[
\Delta C_A = -\frac{1}{1 + x_o} \Delta C_T \tag{15.15}
\]

Now \(C_T^{eq}\) is a function of \(C_a\) as described by Equation 15.7, such that:

\[
C_T^{eq} \approx C_T^{eq}(0) + \frac{dC_T^{eq}}{dC_A} \Delta C_A \tag{15.16}
\]

Substituting Equations 15.15 and 15.16 into 15.12 yields an Equation for the perturbation to the land carbon:

\[
\frac{d\Delta C_T}{dt} = \frac{-\Delta C_T}{\tau} \left[ 1 + \frac{1}{(1 + x_o) \frac{dC_T^{eq}}{dC_A}} \right] \tag{15.17}
\]

This is a linear Equation with a solution of the form \(\Delta C_T = Ke^{\lambda t}\) where \(\lambda\) is the growth-rate of the linear instability,

\[
\lambda = \frac{1}{\tau} \left[ 1 + \frac{1}{(1 + x_o) \frac{dC_T^{eq}}{dC_A}} \right] \tag{15.18}
\]

The condition for linear instability or 'runaway' is \(\lambda > 0\), i.e.,

\[
\frac{dC_T^{eq}}{dC_A} < -\left(1 + x_o\right) \tag{15.19}
\]

This is much more stringent than the condition for positive feedback (\(dC_T^{eq}/dC_A < 0\)).

Equations 15.7, 15.8 and 15.19 together provide a condition for runaway in terms of the CO₂ concentration (\(C_a\)) and parameters associated with the climate change and the carbon cycle response (\(\mu_s, C_{a0.5}, \Pi_{max}, \phi_c, \frac{df}{dT}\)).

Now we search for the conditions necessary for runaway to occur at any CO₂ concentration, by determining whether the minimum value of \(\frac{dC_T^{eq}}{dC_A}\) satisfies Equation 15.19. The minimum value occurs where \(\frac{d^2C_T^{eq}}{dC_A^2} = 0\), so we first differentiate Equation 15.8 with respect to \(C_a\):

\[
\frac{d^2C_T^{eq}}{dC_A^2} = \frac{C_T^{eq}}{C_a^2(C_a + C_{a0.5})} \left[ \mu_s (1 + 1) C_a^2 \right.
\]

\[
+ 2(\mu_s^2 - 2)C_{a0.5} C_a + \mu_s (\mu_s - 1) C_{a0.5} \]

\(\tag{15.20}
\]

The turning points of \(\frac{dC_T^{eq}}{dC_A}\) occur where the quadratic equation within the square brackets is zero. The root corresponding to the minimum value (i.e., maximum positive feedback) is given by:

\[
C_a = \frac{(1 - \mu_s)}{\mu_s} \left[ 1 + \frac{1}{\sqrt{1 - \mu_s^2}} \right] \tag{15.21}
\]

Equation 15.21 gives the critical CO₂ concentration at which the positive feedback from the carbon cycle is strongest. Note that this critical CO₂ concentration is always larger than the critical CO₂ concentration for sink-to-source transition (see Figure 15.3).

![Figure 15.3](image-url) The critical CO₂ concentrations beyond which the land becomes an overall source of CO₂ (dashed line), and at which the positive feedback is maximised (continuous line), as a function of the control parameter, \(\mu_s\). These curves are calculated from Equations 15.10 and 15.21 respectively.
By substituting Equation 15.21 into Equation 15.8, we can determine the most negative value of $\frac{\partial C_T}{\partial C_a}$ which represents the strongest positive feedback from the land carbon cycle (Figure 15.4). Only values which satisfy Equation 15.19 are capable of producing a runaway feedback/linear instability, which requires $\frac{\partial C_T}{\partial C_a} < -1$ even in the absence of ocean carbon uptake (i.e., $\chi_o = 0$). This necessary condition for a runaway land carbon cycle feedback is represented by the horizontal dashed line in Figure 15.4. Note that $\mu > 0.9$ is required for runaway.

The fact that such a large value of $\mu$ implies a present-day land carbon source (Figure 15.1), indicates that a land carbon cycle runaway in the future is unlikely given the existence of a current-day land carbon sink. Figure 15.5 shows the separation of the ‘Current-day Carbon Sink’ and ‘Runaway Feedback’ regions in the $(\Delta T_{2^\circ C}, q_{10})$ parameter space.

15.4 Conclusions

The results from offline dynamic global vegetation models (Cramer et al. 2001) and from the first generation coupled climate-carbon cycle GCMs (Friedlingstein et al. 2003), suggest that climate change will adversely affect land carbon uptake. In some models this effect is strong enough to convert the current land carbon sink to a source under 21st century climate change (Cox et al. 2000). In this paper we have applied a very simple land carbon balance model to produce an analytical expression for the critical CO$_2$ concentration at which the source-to-sink transition will occur. Beyond this critical point the land carbon cycle accelerates anthropogenic climate change, so this also represents one possible definition of ‘dangerous climate change’ in the context of the United Nations Framework Convention on Climate Change.

We have found that the critical CO$_2$ concentration for such a sink-to-source transition in the land carbon cycle is dependent on a single control parameter ($\mu$), which is itself dependent on the climate sensitivity to CO$_2$ and the sensitivities of photosynthesis and ecosystem respiration to climate. Relatively small changes in these parameters can change the critical CO$_2$ concentration significantly, helping to explain why most existing terrestrial carbon cycle models do not produce a sink-to-source transition in the 21st century.

We have also used the simple carbon balance model to examine the necessary conditions for a runaway land carbon cycle feedback. A runaway occurs when the gain factor of the (climate-land carbon storage) feedback loop exceeds one, which is equivalent to the condition for the system to be linearly unstable to an arbitrary perturbation. In this case a change in atmospheric CO$_2$ concentration could occur in the absence of significant anthropogenic emissions, leading to a rapid climate change (i.e. one that is potentially much faster than the anthropogenic forcing that prompted it). We have shown that the condition for such a runaway feedback is much more stringent than the condition for a positive feedback. Furthermore, although a runaway is theoretically possible (e.g. if the climate sensitivity to CO$_2$ is very high), the simple model indicates that such a strong land carbon source in the future is unlikely given the existence of a land carbon sink now.

Our analysis confirms the importance of reducing the uncertainties in eco-physiological responses to climate change and CO$_2$ if we are to be forewarned of a possible source-to-sink transition in the land carbon cycle. However,
it also highlights the critical nature of uncertainties in the climate sensitivity, which not only determines the magnitude of climate change for a given CO₂, but also influences the strength of the land carbon cycle feedback, and therefore the anthropogenic emissions consistent with stabilisation at the given CO₂ concentration (Jones et al., this volume).

Acknowledgements

This work PMC was supported by the European Commission under the ‘CAMELS’ project (PMC and CH); the UK Department of the Environment, Food and Regional Affairs, under contract PECG 7/12/37 (CDJ and PMC); and Science Budget funding from the Centre for Ecology and Hydrology (CH).

REFERENCES

Jones, C.D., P.M. Cox and C. Huntingford, 2005: Impact of climate-carbon cycle feedbacks on emissions scenarios to achieve stabilisation. This volume.