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Balancing the carbon budget. Implications for projections of future carbon dioxide concentration changes

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ABSTRACT

A carbon cycle model is described incorporating CO_2 fertilization feedback and a convolution ocean model that allows the atmosphere-to-ocean flux to be varied. The main parameters controlling the model's behaviour are a fertilization feedback parameter (r) and an ocean flux scaling factor (characterized by the mean carbon flux into the ocean over the 1980s, F(1980s)). Since the model's 1980s-mean net land-use-change flux (D_n(1980s)) is a unique function of r and F(1980s), the model's behaviour can also be characterized by specifying D_n(1980s) (instead of r) and F(1980s). The history of past land-use fluxes, $D_n(t)$, is derived by inverse modelling for a range of values of F(1980s) (1.0–3.0 GtC/yr) and D_n(1980s) (0.6–2.6 GtC/yr). Even with this flexibility, the resultant $D_n(t)$ differs markedly from the observationally-based record of Houghton, particularly before 1950. The inverse calculations are used to determine the history of the so-called "missing sink", as implied directly by the model and by the observationally-based record of $D_n(t)$, for a range of ocean uptake efficiencies as defined by F(1980s). Projections of future CO_2 concentration changes are made for the 6 emissions scenarios recently produced by the Intergovernmental Panel on Climate Change (IS92a–f). The ability to specify F(1980s) and D_n(1980s) allows one to account for the missing sink in a variety of ways, and to account for uncertainties in the amount of missing carbon. This leads to a range of projections and provides some insights into the uncertainties surrounding these projections.

1. Introduction

The basic equation for determining changes in atmospheric carbon dioxide mass (M (GtC)) or concentration (C (ppmv)) is:

$$dM/dt = 2.123\, dC/dt$$

$$= \sum (\text{sources}) - \sum (\text{sinks}) \qquad (1)$$

If one considers only the main sources (industrial emissions and fluxes related to land-use changes) and assumes that the ocean is the only sink, then the budget expressed by eq. (1) appears not to balance. This problem is exemplified using 1980s-mean data from the Intergovernmental Panel on Climate Change (IPCC; Watson et al., 1990): 5.4 ± 0.5 GtC/yr for industrial emissions (IPCC gave the 1987 value of 5.7 GtC/yr) and 1.6 ± 1.0 GtC/yr for emissions due to land-use changes. Since the current rate of atmospheric concentration build-up has been equivalent to around 3.4 GtC/yr over the 1980s (uncertainty ± 0.2 GtC/yr), the residual sink averaged over this decade must be 3.7 ± 1.7 GtC/yr.

The central value of this range is much larger than the IPCC estimate of the mean ocean flux over the 1980s, 2.0 ± 1.0 GtC/yr (Watson et al., 1990, p. 17), implying the existence of an additional (i.e., "missing") sink. If the ocean flux were as low as 1 GtC/yr, as suggested by Tans et al. (1990), then the missing sink could have a 1980s-mean value as large as 4 GtC/yr. Such an extreme value is highly unlikely, however, since recent work has supported the IPCC ocean flux range (Sarmiento and Sundquist, 1992; Robertson and

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Watson, 1992). Using 2.0 ± 1.0 GtC/yr, as the 1980s-mean ocean flux, yields a missing sink of 1.7 ± 2.7 GtC/yr*.

The most immediate question arising from this discrepancy is: where and how large is the missing sink? A secondary, but equally important issue is: what does the existence of a non-oceanic sink mean for projections of future atmospheric CO_2 increases? More generally, how sensitive are such projections to carbon cycle model uncertainties? These are the questions addressed in this paper.

2. Carbon cycle modelling

In the following, a simple and highly idealized carbon cycle model will be developed specifically to address the questions asked above. The main requirements of the model are that it should be flexible enough to consider a range of possible atmosphere-to-ocean fluxes, and that it should include the main terrestrial sources, sinks and feedbacks. Before describing the model, some general issues will be discussed.

Carbon cycle models differ in many aspects, not least in the efficiency with which CO_2 is transferred from the atmosphere to the ocean. Typical box-diffusion models give an ocean flux of 2.0–2.5 GtC/yr for the 1980s (e.g., 2.4 GtC/yr; Siegenthaler, 1983), while ocean general circulation models with embedded carbonate chemistry give values between 1.2 GtC/yr (Maier-Reimer and Hasselmann, cited by Watson et al., 1990, p. 13) and 1.9 GtC/yr (Maier-Reimer and Hasselmann, 1987; Sarmiento et al., 1992). The highest values have been obtained with an outcrop-diffusion model, viz 3.6 GtC/yr (Siegenthaler, 1983), although this is generally thought to be an overestimate because the model assumes instantaneous exchange between the atmosphere and the deep ocean through outcrop zones. It is on the basis of such information, together with empirical estimates such as those of Keeling et al. (1989) and Tans et al. (1990) that IPCC suggested the range of 2.0 ± 1.0 GtC/yr.

When models such as those cited above are used to make projections of future atmospheric CO_2 concentration changes, they are often tuned first to match past concentration changes. This tuning may be performed, not by changing internal model parameters (which may be set on the basis of, for example, isotopic data such as ^{14}C), but by adjusting the somewhat uncertain history of past net emissions changes to match past concentration changes. This can be done in a straightforward way by inverting the carbon cycle model calculations (e.g., Enting and Pearman, 1987; Siegenthaler and Oeschger, 1987; Wigley, 1991). The implied net emissions clearly depend on the particular model; specifically, on how efficiently the model transfers CO_2 from the atmosphere to the oceans. In the calculations described below, the inversion calculations are generalized by varying the overall magnitude of the ocean flux. The model to be used has a free parameter that can be adjusted so as to match a particular instantaneous ocean flux value (e.g., for a specific year) or integrated flux over a specified time period.

Inverse calculations interpret concentration changes in terms of changes in net emissions, i.e., the sum of industrial emissions (fossil fuel burning, cement production, gas flaring), emissions due to land-use changes (mainly those associated with deforestation), and other unspecified sources and/or sinks. It is the sum of these unspecified sources and/or sinks that corresponds to the missing sink noted earlier. The difference between the net emissions so calculated and independent data for the industrial and land-use-change emissions provides an estimate of changes in the missing sink term.

A number of explanations for the missing sink have been proposed, all of which require the missing carbon to be stored in the terrestrial biosphere. The most obvious possibility is that the land-use-change emissions have been overestimated, due, for example, to inadequate provision for regrowth (Kauppi et al., 1992) and/or to an underestimation of the amount of burnt material transferred to and stored more or less permanently in the soil zone (Goudriaan, 1989). Alternatively, there may have been an enhancement of biomass productivity due to the CO_2 "fertilization" effect, nitrogen fertilization from industrial pollution, and/or climate change.

If a carbon cycle model is to be used to make

* Note that if the uncertainty ranges for the individual carbon budget components represent p% confidence intervals, then their sum, ± 2.7 GtC/yr, will be the P% interval where $P > p$. For comparability, the equivalent p% range is around ± 1.5 GtC/yr (see also Table 1.)

Table 1. *Components of the carbon dioxide mass balance*

Component	1980s-mean values (GtC/yr)	1850–1989 totals[a] (GtC)
2.123 dC/dt or 2.123 ΔC	3.4 ± 0.2	141 ± 10[b]
Ocean flux (F)	2.0 ± 1.0	111 ± 56[c]
Industrial sources (I)	5.5 ± 0.5[d]	213 ± 20[d]
Net land-use changes (D_n)	1.6 ± 1.0	123 ± 40[e]
Biomass changes for balance[f]	-0.1 ± 1.1	39 ± 60
Apparent budget imbalance[g]	-1.7 ± 1.5	-80 ± 72

Except where indicated, figures come from IPCC (Watson et al., 1990). Primary uncertainty limits are from IPCC. For derived quantities, limits have been estimated to correspond to the same confidence interval as for the primary variables.

[a] i.e., 1850–1989 inclusive (140 years).

[b] Based on smoothed synthesis of ice-core data from Neftel et al. (1985) and Friedli et al. (1986), and Mauna Loa measurements (Keeling and Whorf, 1991).

[c] Estimated using the present model. Central value corresponds to a 1980s flux of 2.0 GtC/yr, with the uncertainty bounds corresponding to the 1980s uncertainties of ± 1.0 GtC/yr.

[d] From data of Keeling (1991) and Marland and Boden (1991).

[e] Based on 1850–1986 estimate of 117 ± 35 GtC given by Watson et al. (1990, p. 13).

[f] Net terrestrial biomass change required to balance the atmospheric CO_2 mass increase using the above data for F and I: i.e., $\Delta B = 2.123\,\Delta C + F - I$.

[g] i.e., $2.123\,\Delta C + F - I - D_n$: assumes that these are the only components of the balance. The negative of this number gives the magnitude of the "missing sink".

future concentration projections, then there are two options: either processes representing the missing sink can be built into the model, or the missing sink can be ignored. If one were to consider only the 1980s budget, then ignoring the missing sink may be justifiable since a zero value lies within the range of uncertainties. This is also possible if one considers the integrated budget over 1850–1989 (see Table 1). However, it is not possible to match the full time histories of modelled and empirical net fluxes, even within their wide uncertainty ranges. Even if it were, on the basis of the best estimates of the various components of the carbon budget, it is clearly preferable to build some additional sink mechanism into any model to be used for future CO_2 projections. In other words, the model should, in some way, account for the missing sink.

In the present model, only the CO_2 fertilization effect, which acts as a negative feedback effect on concentration build up, is considered. Numerous other models have been developed that include this mechanism (e.g., Bacastow and Keeling, 1973; Björkström, 1979; Goudriaan, 1989; Harvey, 1989a). In principle, the strength of the fertilization feedback may be adjusted to optimise the balance of the carbon budget. However, even if this is done, the balance may be spurious if the model ocean produces an incorrect atmosphere-to-ocean flux. Thus, it is an additional advantage if the model allows a range of ocean fluxes to be considered.

3. The carbon cycle model

The carbon cycle model described below, while extremely simple, captures many of the essential features of more complex models. It was designed specifically so that the sensitivities of the carbon dioxide system to some of the major areas of uncertainty in carbon cycle modelling could be easily and quickly explored. The model is computationally highly efficient and it takes only a fraction of a second to run a 300-year simulation on a modest (80486-based) microcomputer.

The model considers six carbon reservoirs: the oceans, the atmosphere, and four terrestrial carbon stores. Terrestrial feedbacks due to changes in temperature and atmospheric carbon dioxide concentration are included (although temperature feedbacks are ignored in the present analysis), and the method by which the atmosphere-ocean flux is

modelled allows this time-dependent flux to be adjusted in an internally consistent way. Much more complex models exist, but a compromise has to be reached between the detail and transparency of the model (since the ease with which results may be interpreted tends to be inversely related to the number of potentially tunable model parameters).

The basic equation is the mass balance relationship given by eq. (1), which may be rewritten as

$$2.123\ \mathrm{d}C/\mathrm{d}t = 2.123\ \mathrm{d}\Delta C/\mathrm{d}t = E(t) - F(t). \qquad (2)$$

Here, $\Delta C = C - C_0$ is the concentration change from a pre-industrial initial level of C_0, $E(t)$ is the total (net) emissions and $F(t)$ the atmosphere-to-ocean flux. $E(t)$ and $F(t)$ are both expressed as changes relative to their pre-industrial levels (it is assumed that steady-state conditions apply initially). In general, F should be the sum of all fluxes into the ocean, including carbon from weathering transported by rivers. The absolute value of the weathering flux is certainly significant (Sarmiento and Sundquist, 1992), but its changes are probably small and they are ignored here.

Emissions may be divided into two components, those due to industrial activity (I), mainly fossil fuel burning, and those due to changes in the terrestrial carbon store (viz. $\mathrm{d}B/\mathrm{d}t$). The latter in turn may be divided into the net flux due to land-use changes (D_n), mainly deforestation, and carbon store changes associated with the primary terrestrial feedbacks (X). Hence,

$$E = I - \mathrm{d}B/\mathrm{d}t = I + D_n - X, \qquad (3)$$

so that eq. (2) becomes

$$2.123\ \mathrm{d}\Delta C/\mathrm{d}t = I + D_n - F - X. \qquad (4)$$

The term X represents the missing sink (positive X means a transfer from the atmosphere to the terrestrial biosphere).

To solve eqs. (3) and (4), additional equations determining $\mathrm{d}B/\mathrm{d}t$ and F are required. $\mathrm{d}B/\mathrm{d}t$ is determined by the terrestrial component of the model, which is described in detail in the Appendices. The ocean flux, F, is determined here using the convolution integral representation of the ocean general circulation carbon cycle model of Maier-Reimer and Hasselmann (1987); see also Harvey (1989a). As described by Wigley (1991), the model can be written in the form

$$\mathrm{d}\Delta C/\mathrm{d}t = E(t)/2.123 - \sum_{j=0}^{4} \Delta C_j/\tau_j, \qquad (5)$$

where ΔC_j are partial concentrations defined by

$$\Delta C_j(t) = a_j \int_0^t (E(u)/2.123) \times \exp(-(t-u)/\tau_j)\,\mathrm{d}u. \qquad (6)$$

In eq. (6), a_j and τ_j ($j = 0, 4$) are convolution parameters (the a_j sum to one, and $\tau_0 = \infty$) and the partial concentrations sum to ΔC. The partial concentrations do not have to be calculated by direct integration of eq. (6), which is computationally demanding, but can be obtained much more efficiently from the equivalent differential equation

$$\mathrm{d}\Delta C_j/\mathrm{d}t = a_j E(t)/2.123 - \Delta C_j/\tau_j. \qquad (7)$$

When summed over j, eq. (7) leads to eq. (5). Comparing eqs. (2) and (5) shows that the ocean flux, $F(t)$, is determined by the partial concentrations according to

$$F(t) = 2.123 \sum_{j=0}^{4} \Delta C_j/\tau_j. \qquad (8)$$

By a simple scaling of the lifetimes, therefore (i.e., replacing τ_j by τ_j/ξ), one can alter the ocean flux (see, e.g., Harvey, 1989a, and Wigley, 1991).

Past or future CO_2 concentration changes for any prescribed scenario of industrial emissions, $I(t)$, and land-use-change emissions, $D_n(t)$, can now be obtained by solving simultaneously eq. (7), eq. (3) (which defines $E(t)$ in terms of $I(t)$ and changes in the terrestrial carbon store, $\mathrm{d}B/\mathrm{d}t$), and eqs. (B2), (B3), (B4), (B5) and (B9) (which determine $\mathrm{d}B/\mathrm{d}t$ as a function of $D_n(t)$, see Appendix B). In finite difference form, the equations become a simple set of coupled algebraic equations which can be solved accurately and swiftly with a time step of one year. The model contains a number of parameters whose values must be set; specifically, the convolution constants a_j and lifetimes τ_j (see Wigley, 1991, Table 2), various initial terrestrial fluxes and reservoir sizes, constants that determine how fluxes between the

Table 2. *Model parameter values required to balance the carbon budget for specified values of D_n(1980s) and F(1980s); units GtC or GtC/yr*

D_n(1980s)	F(1980s)	r	β340	$\sum D_n$(1850–1989)
0.6	1.0	1.29	0.52	89
0.6	2.0	1.08	0.11	89
0.6	3.0	0.92	−0.11	83
1.6	1.0	1.54	0.93	142
1.6	2.0	1.26	0.41	146
1.6	3.0	1.06	0.08	144
2.6	1.0	1.91	1.79	192
2.6	2.0	1.51	0.86	200
2.6	3.0	1.23	0.36	202

r is the required value of the CO_2 fertilization coefficient (r = NPP(680 ppmv)/NPP(340 ppmv)) and β340 is the effective "β factor" at 340 ppmv as defined in Appendix A.

terrestrial reservoirs are partitioned, the CO_2 fertilization feedback parameter r or β, and the ocean flux scaling factor ξ. The fertilization effect is modelled here using the rectangular hyperbola representation for which the adjustable parameter is r. r is equivalent to the β factor which appears in the logarithmic formulation. Further details are given in Appendix A. The parameters for and the structure of the terrestrial component of the model are discussed in Appendix B.

4. Explaining past concentration changes

The apparent imbalance of the carbon budget is simply an indication that we are not able to quantify properly all the terms in the budget. In terms of concentration changes, this is manifest when we force a carbon cycle model that has no missing sink term with best-guess estimates of the history of industrial and land-use-change emissions. The resulting predicted concentration values are considerably higher than those observed. If a missing sink term is included in the model, the agreement can be improved, but discrepancies will always remain.

An alternative way to express these discrepancies is to use a carbon cycle model in inverse mode, i.e., to begin with the observed concentration changes and deduce from these the required history of net emissions, $E(t)$. If the model has a deterministic missing sink term, then its time evolution, $X(t)$, will be derived automatically by the model. Adding these two terms $(E(t)+X(t))$ gives the sum of industrial and land-use-change emissions that is required for consistency with the observed concentration changes (see eq. (3)). Thus, if $I(t)$ is known, $D_n(t)$ can be obtained. Comparing this $D_n(t)$ with an observationally-based record (such as that of Houghton, 1991) provides another view of uncertainties in the carbon budget.

The derived history of net land-use-related emissions, $D_n(t)$, depends, of course, on how the model quantifies the main sink mechanisms, $F(t)$ to the ocean and $X(t)$ to the terrestrial biomass. In the present model, these are determined by the ocean-flux scaling factor, ξ, and the fertilization factor, r. (The terrestrial sink is also affected by values of the other model parameters in the land component of the model, see Appendix B, but these are taken as fixed quantities here since most of the variability in the results arises through the range of possible values of the fertilization parameter.) Thus, a range of $D_n(t)$ histories can be derived by choosing different values of ξ and r (within their recognized uncertainty ranges).

Before considering the results of such inverse calculations, the calculation method will be given in a little more detail. The basic problem is to find, for a given value of the fertilization parameter r, the value of $D_{n,i}$ (the net land-use-change emission in year i) such that the concentration change over the year calculated by the model

$$C_i - C_{i-1} = (I_i + D_{n,i} - F_i - X_i)/2.123 \qquad (9)$$

is the same as that observed. This is most efficiently done by iteration, required because X_i depends on $D_{n,i}$. The number of iterations needed is small provided the first estimate of $D_{n,i}$ is a good one.

As noted earlier, the resulting $D_n(t)$ depends on the choice of ξ and r. To characterize the ocean flux term, the mean value of the flux over the 1980s (F(1980s)) can be used instead of ξ, since, in the convolution representation of the ocean component of the model there is a one-to-one correspondence between ξ and F(1980s). The range 2.0 ± 1.0 GtC/yr is used for F(1980s).

For r, an alternative constraint can be applied to define the range of values used. For any choice of values of the pair $F(1980s) - r$, a unique value of D_n(1980s) is determined. Thus, instead of using r

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as a primary specification variable, D_n(1980s) may be used. This has the advantage that D_n(1980s) has a recommended (IPCC) uncertainty range, whereas r does not. In performing the inverse calculations, therefore, a particular pair of values for F(1980s) − D_n(1980s) is chosen initially and an additional set of iterations is performed varying r in order to match the chosen value of D_n(1980s). This iterative procedure can also be carried out very efficiently by making a judicious initial choice for r.

The results of such inverse calculations with the present model are shown in Figs. 1, 2 and Table 2. Three sets of values of F(1980s) (1, 0, 2.0 and 3.0 GtC/yr) and D_n(1980s) (0.6, 1.6 and 2.6 GtC/yr) are used. Figs. 1, 2 also show the history of net land-use-change emissions from Houghton (1991, updated).

Fig. 1 shows $D_n(t)$ for the central value of D_n(1980s) and various values of F(1980s). In both Figs. 1 and 2, the early (pre-1860) short-term fluctuations arise from changes in $\mathrm{d}\Delta C/\mathrm{d}t$, while the later fluctuations arise mainly from short-term fluctuations in the history of industrial emissions.

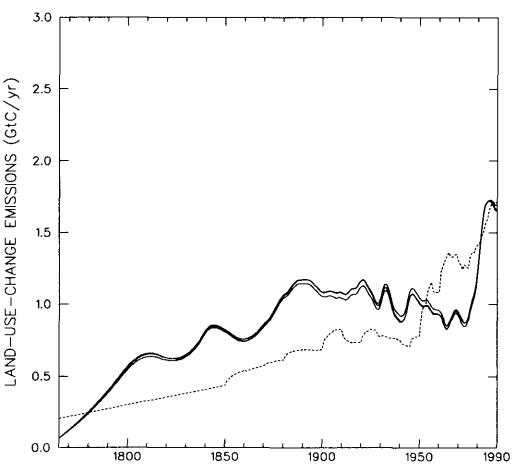

Fig. 1. Land-use-change emissions ($D_n(t)$) required to match past concentration data for different efficiencies of ocean CO_2 uptake (1980s-mean fluxes of 1.0, 2.0 and 3.0 GtC/yr) and with the land-use flux constrained to have a 1980s mean of 1.6 GtC/yr. Concentration data from Neftel et al. (1985), Friedli et al. (1986) and Keeling and Whorf (1991). Industrial emissions data from Keeling (1991) and Marland and Boden (1991), smoothed using a nine-term binomial filter. Observationally-based emissions, from Houghton (1991, updated), are shown dashed.

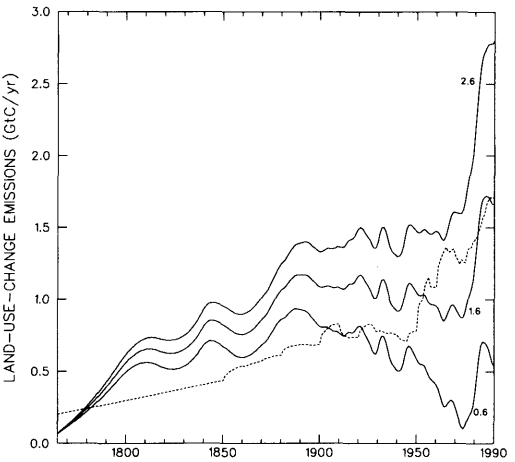

Fig. 2. Inferred land-use-change emissions ($D_n(t)$) for different values of the 1980s-mean flux (D_n(1980s) = 0.6, 1.6 and 2.6 GtC/yr) and 1980s-mean ocean flux of 2.0 GtC/yr. As indicated by Fig. 1, very similar results obtain for other values of F(1980s). Observationally-based emissions, from Houghton (1991, updated), are shown dashed.

Because $\Delta C(t)$ changes are smooth, so too are changes in $X(t)$ and $F(t)$, leaving only $D_n(t)$ in the mass balance to compensate for relatively rapid changes in $I(t)$. Clearly, this is just an artefact of the inverse calculation procedure. The detailed changes in $D_n(t)$ must be independent of $I(t)$, so short time scale variations in $D_n(t)$ derived by inverse means have no real significance.

In Fig. 1, it can be seen that the $D_n(t)$ values are relatively insensitive to F(1980s). This is because a higher value of F(1980s) requires a lower r value to maintain the specified value for D_n 1980s), and the consequent reduced terrestrial sink largely compensates for the increased ocean uptake over the whole time period. (For any fixed value of r, the implied $D_n(t)$ history is, of course, quite sensitive to the choice of F(1980s).)

Fig. 2 shows $D_n(t)$ for the central value of F(1980s) and a range of D_n(1980s) values. The general character of $D_n(t)$ shows a rapid increase in the late 18th century, followed by a period of steadily increasing land-use emissions to around 1890. The early rapid increase in $D_n(t)$ is partly an artefact of uncertainties in the observed concentration changes. Relatively minor changes in the concentration data produce noticeably different $D_n(t)$ histories, simply because inverse modelling is very

sensitive to dC/dt. For $D_n(1980s) = 0.6$ GtC/yr, $D_n(t)$ increases to around 1890 and then declines somewhat erratically to near zero in the mid 1970s, before showing the rapid increase that is common to all $D_n(1980s)$ cases. For $D_n(1980s) = 1.6$ and 2.6 GtC/yr, $D_n(t)$ remains at roughly the same level from 1890 to the mid 1970s. The most striking feature of the recent record is the very rapid rise in $D_n(t)$ in the late 1970s/early 1980s, similar to the Houghton data. The Houghton data, however, begins this rise in 1950, whereas the inverse results do not show a rise until around 1973. Indeed, for $D_n(1980s)$ less than about 2 GtC/yr, the inverse results show a marked decline in land-use emissions over 1945–73, opposite to the Houghton data trend.

Another interesting result from these calculations is the apparent inconsistency between the inverse estimates of $D_n(1980s)$ and the sum of emissions over 1850–1989 ($\sum D_n(1850–1989)$; see Table 2). The IPCC estimates are $D_n(1980s) = 1.6 \pm 1.0$ GtC/yr and $\sum D_n(1850–1989) = 123 \pm 40$ GtC (see Table 1). From the Houghton data used here, the corresponding values are 1.6 GtC/yr and 121 GtC. For $D_n(1980s) = 1.6 \pm 1.0$ GtC/yr, the inverse calculations give $\sum D_n(1850–1989) =$ 85–200 GtC (if r is constrained to lie in the range 1.0–1.5). Although the two $\sum D_n$ ranges overlap, the inverse calculations suggest a higher value than the observationally-based estimate. Fig. 2 shows that this is due largely to higher inverse deforestation estimates in the earlier part of the record, up to around 1950 for $D_n(1980s) = 1.6$ GtC/yr, the case which is the most similar to the Houghton data in recent decades.

5. History of the "missing sink"

An alternative way to view these results is in terms of the implied history of missing sink changes, $X(t)$. This can be derived in two ways, either directly from the model (and therefore based on the inverse estimates of $D_n(t)$), or using model results for the ocean flux consistent with the observed concentration history, and Houghton's estimate for $D_n(t)$. In both cases, $X(t)$ is defined by (cf. eq. (4))

$$X(t) = I(t) + D_n(t) - F(t) - 2.123\, d\Delta C/dt \qquad (10)$$

The results are shown in Figs. 3, 4.

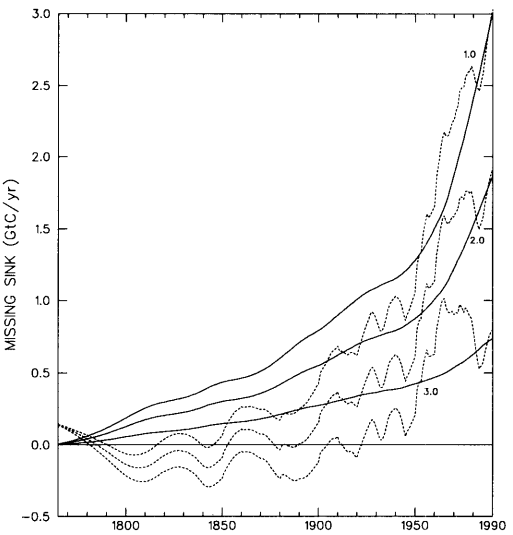

Fig. 3. Missing sink variations for different ocean uptake efficiencies ($F(1980s) = 1.0$, 2.0 and 3.0 GtC/yr). Smooth curves are derived directly from the model, are dependent only on the way CO_2 fertilization is modelled and reflect the smoothly varying CO_2 concentration changes. The more variable curves are the difference (industrial emissions plus net land-use-change emissions from Houghton (1991, updated)) minus (ocean uptake plus atmospheric mass increase).

Fig. 3 compares the inverse and Houghton results for different ocean uptake efficiencies, $F(1980s) = 1.0$, 2.0 and 3.0 GtC/yr. For any given $F(1980s)$ value, the missing sink implied by the inverse calculation is larger than the value implied by Houghton's data up to around 1950. As noted earlier, this is because the inverse values of $D_n(t)$ are greater than Houghton's for all except the most recent decades. The inverse values vary smoothly because they reflect the smoothly varying concentration changes that determine the missing sink in the model. The Houghton values vary on shorter time scales, reflecting similar time scale variations in Houghton's $D_n(t)$ and in the industrial emissions term, $I(t)$. The real-world missing sink should also show decadal and shorter time scale variability, reflecting climate-related variations in terrestrial processes and ocean uptake. The changes evident in Fig. 3, however, cannot be interpreted as such; more likely they reflect uncertainties in $D_n(t)$ and $I(t)$. The near-zero values of the Houghton-based missing sink up to around

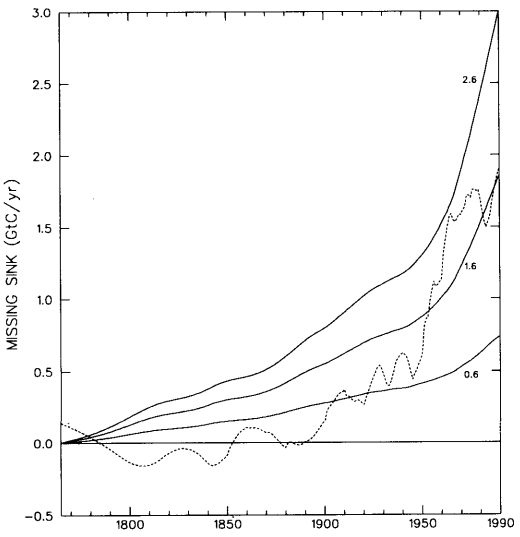

Fig. 4. Missing sink variations for different values of the 1980s-mean net land-use-change emissions (D_n(1980s) = 0.6, 1.6 and 2.6 GtC/yr) and ocean uptake F(1980s) = 2.0 GtC/yr. Smooth curves are derived directly from the model (cf. Fig. 3). The more variable curve is based on land-use-change emissions from Houghton (1991, updated).

1890, the rapidity of the rise over 1890 to 1955 (more rapid than any of the inverse curves), and the levelling off of the sink subsequently should not be over-interpreted. They could as well reflect data input uncertainties as physically meaningful variations in the sink.

In Fig. 3, the implied history of the missing sink was shown to be highly sensitive to the efficiency of ocean uptake. A larger ocean sink obviously implies a smaller missing sink. Fig. 4 shows the sensitivity of the missing sink to uncertainties in the current level of land-use-related emissions for a 1980s ocean flux of 2.0 GtC/yr. In the present model, larger values of D_n(1980s) require a larger CO_2 fertilization effect (i.e., larger r), which in turn leads to a larger missing sink. The Houghton-based missing sink history for F(1980s) = 2.0 GtC/yr is shown for comparison. Since D_n(1980s) for Houghton's data is fixed (and close to 1.6 GtC/yr) it is not possible to estimate D_n(1980s)-related uncertainties for the Houghton-based missing sink.

6. Future concentration changes

The most recent projections of future CO_2 concentrations made by IPCC are those given in the 1990 report (Houghton et al., 1990). Although IPCC gave only a single concentration projection for each of the four emissions scenarios considered, this was based on results from a number of carbon cycle models. In all but one of the models used, terrestrial feedback processes were ignored. The various simulations began with a somewhat unrealistic emissions starting point of around 6 GtC/yr in 1990, compared with about 8 GtC/yr for the sum of industrial and land-use-change emissions. Essentially, the missing sink term was accepted in choosing the starting point (if one takes 6 GtC/yr to be a net emissions value), but ignored in carrying out the projections.

Clearly, a model that gives a realistic carbon balance today and in the past must give more defensible future projections than one which does not. Just how important is it to ensure a realistically balanced starting point for projections beyond 1990? At least qualitatively, the answer to this question is obvious. If negative feedbacks are required to balance the carbon budget today, then ignoring these in the future will lead to an overestimate of future concentration levels. The main argument in favour of the present balanced-model approach is the strong evidence for a CO_2 fertilization effect, both at the small scale (e.g., Gates, 1985; Shugart et al., 1986) and at the global scale (e.g., Gifford, 1980; Keeling et al., 1989; Enting and Mansbridge, 1991).

In the following, future projections will be made under the constraint of a realistic carbon budget balance. This still leaves open the possibility of a range of projections for any given emissions scenario, since there are quite wide uncertainties regarding the balance. Furthermore, any prescribed balance can be achieved in different ways and these may lead to different future concentrations.

Just what is meant by a model that balances the carbon budget? Clearly, any model must give a balance between the rate of increase of concentration and the sources and sinks that the model considers. The balance issue refers to whether the model budget agrees with observations. In other words, we refer somewhat loosely to a "balanced" model as one that leads to no inconsistencies

Table 3. *1992 IPCC CO_2 emissions scenarios (from Leggett et al., 1992)*

Scenario	IS92a		IS92b		IS92c		IS92d		IS92e		IS92f	
year	D_n	I	D_n	I	D_n	I	D_n	I	D_n	I	D_n	I
1990	1.30	6.10	1.30	6.10	1.30	6.10	1.20	6.10	1.30	6.10	1.30	6.10
1995	1.30	6.60	1.30	6.60	1.30	5.90	1.05	6.25	1.30	6.90	1.30	6.70
2000	1.30	7.10	1.30	6.90	1.30	6.20	0.90	6.60	1.30	7.80	1.30	7.50
2005	1.26	7.94	1.26	7.54	1.26	6.54	0.78	7.02	1.26	8.94	1.30	8.40
2010	1.22	8.68	1.22	8.18	1.22	6.88	0.66	7.54	1.22	10.18	1.30	9.50
2015	1.18	9.42	1.18	9.02	1.18	7.12	0.54	7.96	1.18	11.42	1.30	10.60
2020	1.14	10.26	1.14	9.76	1.14	7.36	0.42	8.38	1.14	12.56	1.30	11.80
2025	1.10	11.10	1.10	10.70	1.10	7.70	0.30	9.00	1.10	14.00	1.30	13.10
2050	0.80	13.70	0.80	13.00	0.70	6.80	0.10	8.90	0.80	19.30	0.90	16.30
2075	0.35	15.95	0.35	15.05	0.25	5.35	0.00	9.30	0.35	26.65	0.45	20.75
2100	−0.10	20.40	−0.10	19.20	−0.20	4.80	−0.10	10.40	−0.10	35.90	0.00	26.60

D_n is emissions from land-use changes and I is from industrial activity (GtC/yr).

between model output and observations, within recognized and acceptable observational and model uncertainties. Such consistency may be achieved in different ways. For example, if the model were run in forward mode forced by land-use and industrial emissions changes that were within their respective uncertainty ranges, then the implied ocean uptake and atmospheric concentration changes would also have to be within their accepted uncertainty bounds. Alternatively, as done here, if the model is run in inverse mode with given industrial emissions and concentration changes, then the implied land-use-change emissions and ocean uptake histories would have to be within the accepted uncertainty ranges for these quantities.

The procedure used to balance the present model has been described in the previous section. The model is constrained to have 1980s-mean atmosphere-to-ocean and land-use-change fluxes within the IPCC-recommended ranges of 1.0–3.0 GtC/yr and 0.6–2.6 GtC/yr, respectively. For any specified $F(1980s) - D_n(1980s)$ pair, the fertilization feedback parameter r is calculated using the inverse modelling procedure described above. This ensures a perfect match between past modelled and observed CO_2 concentration changes.

For future emissions, the 1992 IPCC scenarios are used (from Leggett et al., 1992). Values were linearly interpolated between those given in Table 3. The results for the central emissions scenario (IS92a) are shown in detail in Figs. 5, 6. Fig. 5 shows concentration projections for the IPCC best-guess value of $D_n(1980s)$, 1.6 GtC/yr, and for $F(1980s) = 1.0$, 2.0 and 3.0 GtC/yr*. As noted earlier, the results are relatively insensitive to $F(1980s)$ because the different r values required to give $D_n(1980s) = 1.6$ GtC/yr lead to changes in the terrestrial sink that largely offset those in the ocean sink.

In Fig. 6, concentration projections are shown for $F(1980s) = 2.0$ GtC/yr and different values of $D_n(1980s)$. Fig. 6 also shows a "no-feedback" or "passive biomass" projection for $F(1980s) =$ 2.0 GtC/yr, corresponding to the method used in the 1990 IPCC projections. The differences between the projections in Fig. 6 arise because of different amounts of biomass change. For example, comparing the passive biomass case with the full model result for $D_n(1980s) = 1.6$ GtC/yr shows a concentration difference of 115 ppmv in 2100 implying an additional sequestering of 244 GtC in the biomass.

For the feedback case with $D_n(1980s) =$ 1.6 GtC/yr, the biomass increases by 264 GtC over

* There is a minor inconsistency between the choice of $D_n(1980s)$ used to calculate r and the scenario value used for $D_n(t)$ in 1990 (see Table 3). For $D_n(1980s) =$ 1.6 GtC/yr, the model gives $D_n(1990) \approx 1.6$ GtC/yr, noticeably higher than the IPCC scenario value of 1.3 GtC/yr. In the calculations, $D_n(t)$ was assumed to relax from the 1990 value determined by the inverse calculation to the scenario value over the period 1990–2000.

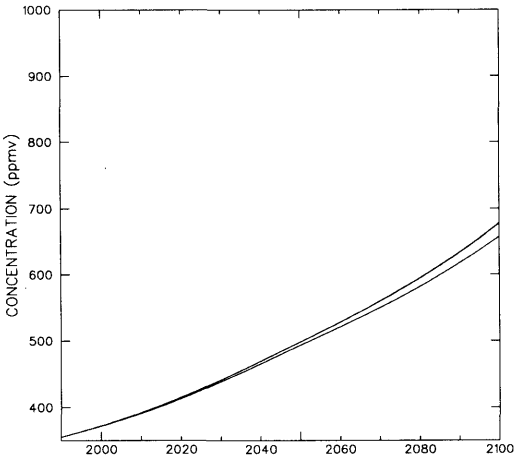

Fig. 5. Projected CO_2 concentration changes for IPCC emissions scenario IS92a for $D_n(1980s) = 1.6$ GtC/yr and $F(1980s) = 1.0$, 2.0 and 3.0 GtC/yr. Higher ocean flux requires a smaller missing sink and feedback parameter to balance the carbon budget over the 1980s, and the two effects largely compensate out to 2100. The $F(1980s) = 1.0$ GtC/yr curve is slightly lower than the other two which are virtually indistinguishable.

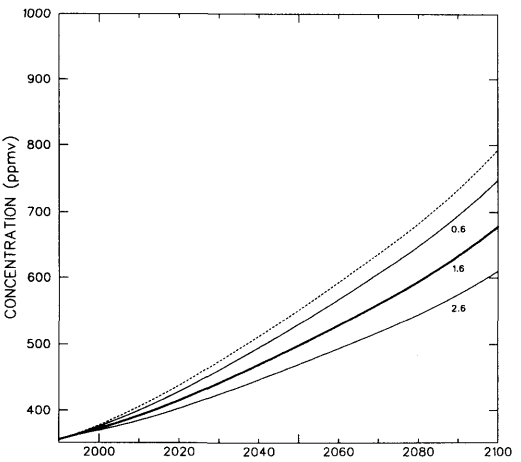

Fig. 6. Projected CO_2 concentration changes for IPCC emissions scenario IS92a for $F(1980s) = 2.0$ GtC/yr. The central (bold) curve is for $D_n(1980s) = 1.6$ GtC/yr (the IPCC best-guess value). The upper two curves are the passive biosphere/no-feedback result (dashed) and the result for $D_n(1980s) = 0.6$ GtC/yr. The lowest curve is for $D_n(1980s) = 2.6$ GtC/yr. *Note*: if the true D_n is $D_{n,\mathrm{obs}}$, then these results correspond to assuming only $I + D_n - F - 2.123\Delta C$ of the missing sink is CO_2 fertilization, with the remainder $(D_{n,\mathrm{obs}} - D_n)$ declining to zero over 1990–2000.

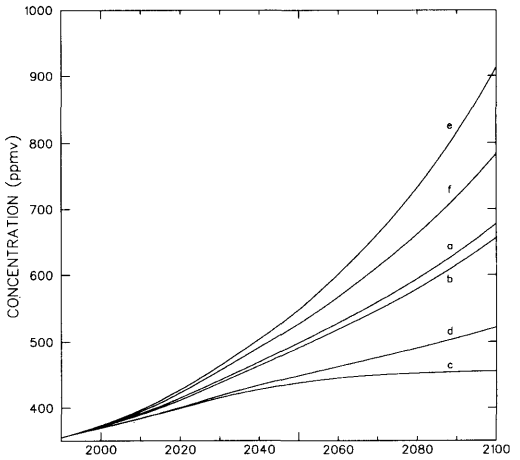

Fig. 7. Best-guess concentration projections for the 1992 IPCC emissions scenarios IS92a–f based on $F(1980s) = 2.0$ GtC/yr and $D_n(1980s) = 1.6$ GtC/yr. Results are insensitive to the choice of $F(1980s)$; cf. Fig. 5.

1990–2100, compared with a cumulative net land-use-related loss of 85 GtC. Of this differential of 349 GtC, approximately 7% accumulates in the litter reservoir, 35% in the soil (dead) carbon reservoir and 58% goes into the living biomass. Given that the CO_2 concentration in the year 2100 is well over double the pre-industrial level, enough to cause large changes in plant biomass produc-

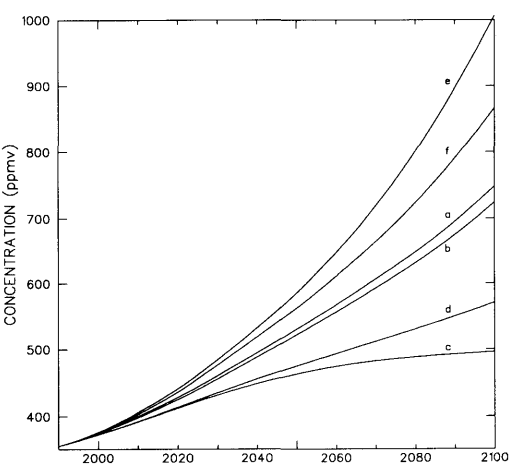

Fig. 8. Upper bound concentration projections for the 1992 IPCC emissions scenarios IS92a–f based on $F(1980s) = 2.0$ GtC/yr and $D_n(1990) = 0.6$ GtC/yr. Results are insensitive to the choice of $F(1980s)$; cf. Fig. 5.

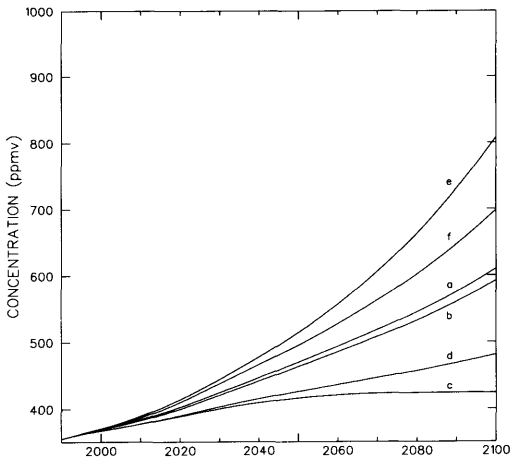

Fig. 9. Lower bound concentration projections for the 1992 IPCC emissions scenarios IS92a–f based on $F(1980s) = 2.0$ GtC/yr and $D_n(1980s) = 2.6$ GtC/yr. Results are insensitive to the choice of $F(1980s)$; cf. Fig. 5.

tivity in small-scale experiments, these amounts seem not unreasonable, although they are, of course, model dependent.

Figs. 7–9 and Table 4 summarize results for the other emissions scenarios. In Fig. 7, the IPCC best-guess value of $D_n(1980s) = 1.6$ GtC/yr is assumed, while Figs. 8, 9 show results for $D_n(1980s) = 0.6$ and 2.6 GtC/yr respectively. The Fig. 7 results differ slightly from those given earlier in Wigley and Raper (1992) for a number of reasons: because of improvements made in the terrestrial component of the model; because the 1980s ocean flux value used here is 2.0 GtC/yr rather than 1.95 GtC/yr used earlier; because 1.6 GtC/yr is used as the best guess value for the 1980s-mean land-use-change emissions (1.1 GtC/yr previously); because the earlier calculations assumed 1980s-mean gross land-use-change fluxes rather than net fluxes as more correctly used here; and because a different functional form is now used for the CO_2 fertilization effect. The combined effect of all these factors is to produce lower concentration projections.

Because of the character of the emissions scenarios, and because of lags in the response of the carbon system to emissions changes, inter-scenario differences are relatively small out to around 2050. After that the results diverge rapidly and the 2100 concentrations span very wide ranges. Since all of these scenarios are based on the general assumption of "existing policies", these ranges reflect basic uncertainties in predicting future population growth and changes in energy use in the absence of strong controls to reduce CO_2 emissions.

Table 4 gives uncertainty ranges for $C(2100)$, and expresses the 2100 results in terms of the

Table 4. *Concentration and radiative forcing change uncertainties*

Projection	LOW[a]		MID[b]		HIGH[c]		NFB[d]		$\delta\Delta Q$[e]
Scenario	$C(2100)$	ΔQ	$C(2100)$	ΔQ	$C(2100)$	ΔQ	$C(2100)$	ΔQ	high–low
IS92a	610	3.41	678	4.07	749	4.70	793	5.06	1.29
IS92b	593	3.23	656	3.87	724	4.49	767	4.85	1.26
IS92c	424	1.12	455	1.57	496	2.10	527	2.49	0.98
IS92d	481	1.91	522	2.43	571	3.00	605	3.35	1.09
IS92e	808	5.18	913	5.95	1006	6.56	1056	6.87	1.38
IS92f	698	4.26	784	4.99	867	5.62	917	5.97	1.36

$C(2100)$ is the concentration level at the end of the year 2100 (ppmv). ΔQ is the radiative forcing change over 1990–2100 (W/m^2).

[a] Using $F(1980s) = 2.0$ GtC/yr and $D_n(1980s) = 2.6$ GtC/yr. The corresponding fertilization factor ($r = 1.51$, see Table 2) is higher than considered likely, so these values probably underestimate the lower limit.

[b] Using $F(1980s) = 2.0$ GtC/yr and $D_n(1980s) = 1.6$ GtC/yr. The corresponding fertilization factor is $r = 1.26$ (see Table 2).

[c] Using $F(1980s) = 2.0$ GtC/yr and $D_n(1980s) = 0.6$ GtC/yr. The corresponding fertilization factor is $r = 1.08$ (see Table 2).

[d] No-feedback/passive biosphere version of the model. Results correspond to those that would obtain from the IPCC90 methodology and probably represent an extreme upper bound.

[e] Radiative forcing difference (W/m^2) in the year 2100 between high and low concentration estimates.

radiative forcing change from 1990 (viz. $\Delta Q = 6.3 \ln(C(2100)/C(1990))$; see Shine et al., 1990). In terms of radiative forcing changes over 1990–2100, the difference between the lowest and highest emissions scenarios is around 4.3 W/m^2 ($\pm$0.2 W/m^2, depending on the values assumed for F(1980s) and D_n(1980s)). The corresponding forcing range between the low and high concentration projections is 1.0–1.4 W/m^2, depending on scenario. While the inter-scenario uncertainty exceeds that due to the carbon cycle uncertainties considered here by three to four times in the year 2100, the relative importance of these uncertainties is strongly time dependent. Out to around 2020, carbon cycle uncertainties in radiative forcing exceed those due to inter-scenario differences.

7. Conclusions

The preceding sections describe a flexible and computationally efficient carbon cycle model that includes CO_2 fertilization feedback and allows the atmosphere-to-ocean flux to be changed. The model has been used to study the apparent imbalance in the carbon budget and to determine the model parameter values required to obtain a realistic balance. Balance is defined as being achieved if the time history of land-use-change emissions obtained by inverse modelling gives values of D_n(1980s) within the range of uncertainty of observationally-based estimates for this quantity.

When the carbon cycle model is run in inverse mode, even given its flexibility with regard to the choice of the efficiency of ocean CO_2 uptake (through F(1980s)) and land-use-change fluxes (through D_n(1980s)), the implied history of past land-use emissions up to 1950 differs markedly from the observationally-based record of Houghton (1991) with the latter being substantially smaller. For D_n(1980s) below about 2.0 GtC/yr, the modelled $D_n(t)$ series declines noticeably over 1950–75, in contrast to the Houghton data. After 1975, both modelled and observationally-based land-use emissions increase rapidly to the mid 1980s, declining slightly or leveling off thereafter.

A further manifestation of these results is that modelled values of $\sum D_n$(1850–1989) appear to be inconsistent with the D_n(1980s) values when compared to Houghton's estimates of these quantities. The differences between the inverse results and Houghton's data could reflect either model deficiencies, and/or uncertainties in the Houghton data, the input concentrations or industrial emissions. The neglect of natural processes (such as climate variations) that could alter terrestrial emissions or ocean uptake may account for some of the decadal time scale differences, but not the consistent longer-term discrepancies.

After balancing the contemporary carbon budget (i.e., ensuring that the model's budget breakdown over the 1980s is consistent with observational evidence), projections of future concentration changes were made for the 1992 IPCC emissions scenarios. A range of projections was obtained for each scenario by considering different values for F(1980s) and D_n(1980s), within their current uncertainty ranges. This is not the first time that some form of budget balancing has been achieved by model tuning prior to making CO_2 concentration projections (see, e.g., Bacastow and Keeling, 1973; Harvey, 1989a). However, it is the first time that inverse calculations spanning a wide range of ocean and land-use-change flux uncertainties, and matching past concentration changes precisely, have been carried out as a precursor to deriving a comprehensive range of future concentration possibilities.

The concentration projections for any given emissions scenario show a wide range of uncertainty. While these uncertainties are related indirectly to model parameter uncertainties, they arise here through current uncertainties in the carbon balance, as epitomized by the magnitude of the missing sink. The primary control on the range of uncertainty is D_n(1980s), the emissions due to land-use changes. This sensitivity of projections to the assumed magnitude of past land-use-change emissions has been noted previously by Harvey (1989a). The more general dependence on the magnitude of the missing sink highlights the importance of obtaining a better explanation for the processes that explain and control this sink.

Concentration projections must depend, not only on the current and past magnitude of the missing sink, but also on its future changes. By accounting for the sink using CO_2 fertilization alone, the present model constrains the form of these future changes. If other processes are important, these may change differently in the future.

Nevertheless, the range of projections produced corresponds to a wide range of changes in the magnitude of the sink with time.

In the high concentration projections, the integrated amount of carbon sequestered by the missing sink over 1990–2100 is 108–149 GtC (the value depends on the scenario). The high projections correspond to a missing sink of 0.68 GtC/yr averaged over the 1980s (see Fig. 4) equivalent to 75 GtC over 110 years. These projections therefore correspond to a slight to moderate increase in the magnitude of the sink over the next century. At the other extreme, the low concentration projections correspond to sequestration of 370–705 GtC over 1990–2100 (although it should be noted that these results correspond to an unlikely high value of the fertilization effect; see Table 2). The mean sink strength over the 1980s for these projections is 2.68 GtC/yr (i.e., 295 GtC over 110 years), so the proportional increase in the magnitude of the missing sink is similar to that for the high concentration projections. These results are consistent with the substantial concentration increases over 1990–2100 and the assumption that CO_2 fertilization is the primary explanation for the missing sink.

In conclusion, the simplicity of the model used here must be stressed. Potentially important processes, such as changes in terrestrial fluxes associated with both past and future changes in climate, have not been considered. These effects could either reduce (Harvey, 1989b; Prentice and Fung, 1990; Smith et al., 1992) or increase (Oechel et al., 1993; Smith and Shugart, 1993) future concentrations over those calculated here. They represent a significant additional source of uncertainty beyond that considered here.

Large uncertainties remain in carbon cycle modelling, and these can only be reduced by more detailed modelling studies supported by observational and process-oriented research. Of prime importance, however, is the need to reduce uncertainties in all aspects of the land-use component of CO_2 emissions.

8. Acknowledgments

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9. Appendix A

Formulation for CO_2 fertilization

The two most commonly used forms for the CO_2 fertilization effect are the logarithmic form introduced by Keeling and Bacastow (1973) and the rectangular hyperbolic or Michaelis-Menton form (Gates, 1985). The former expresses the fertilization enhancement of productivity in terms of a single "β factor" via the expression

$$N = N_0(1 + \beta \ln(C/C_0)), \tag{A1}$$

where N is net primary productivity, C is concentration and subscript zero denotes an initial or reference value. Eq. (A1) behaves unrealistically at both low and high concentrations. In particular, it allows NPP to increase without limit as C increases. The hyperbolic form is

$$N = \frac{(C - C_b)}{(1 + b(C - C_b))} = \frac{N_0(C - C_b)(1 + b(C_0 - C_b))}{[(C_0 - C_b)(1 + b(C - C_b))]}. \tag{A2}$$

Eq. (A2) behaves more realistically in that it leads to zero NPP at $C = C_b$ and has a limiting value as C tends to infinity of

$$N_\infty = N_0(1 + 1/(b(C_0 - C_b))). \tag{A3}$$

I employ eq. (A2), as given by Gifford (1993), and take N_0 and C_0 to correspond to pre-industrial conditions. Following Gifford, I use $C_b =$ 31 ppmv; other authors have used higher values, but the results presented in this paper are insensitive to the precise value of C_b that is used. To determine the other parameter in eq. (A2), b, Gifford uses the enhancement in NPP that occurs for a PCO_2 increase from 340 ppmv to 680 ppmv

$$r = N(680)/N(340). \tag{A4}$$

He suggests that a realistic range of values for r is 1.1–1.4. From eq. (A2), r is related to b by

$$b = \frac{[(680 - C_b) - r(340 - C_b)]}{[(r-1)(680 - C_b)(340 - C_b)]}. \quad \text{(A5)}$$

Specifying r, therefore, determines N/N_0 for given C in the same way that N/N_0 is determined by β in the logarithmic form.

The main distinction between eqs. (A1) and (A2) lies in their behaviour for large C (>1000 ppmv). For any given r value, a β value can be chosen so that the two formulae behave quite similarly over a wide range of concentrations. Nevertheless, when the β and/or r values are obtained by tuning the model to past data, the two fertilization formulations give noticeably different results even for concentrations substantially less than 1000 ppmv.

A convenient way to compare the two formulations is through an equivalent β value (β^*), defined by

$$\beta^* = \partial(N/N_0)/\partial(\ln C). \quad \text{(A6)}$$

Clearly, β^* is constant (independent of C) for the logarithmic formula. For the hyperbolic formula

$$\beta^* = \frac{[(1 + b(C_0 - C_b))\, C]}{[(C_0 - C_b)(1 + b(C - C_b))^2]}, \quad \text{(A7)}$$

which tends to zero for large C. For larger and larger C, the hyperbolic formula therefore gives smaller and smaller enhancement of NPP relative to the logarithmic formula. In describing the results of inverse calculations in which r is adjusted to obtain a specified value for the mean net deforestation flux over the 1980s, to facilitate comparison with the standard β factor of the logarithmic formulation, I give both the r value, and the value of β^* at 340 ppmv calculated using eq. (A7) and taking C_0 to be the pre-industrial concentration (≈ 278 ppmv).

10. Appendix B

The terrestrial component of the carbon cycle model

The terrestrial component of the carbon cycle model is essentially one of the hierarchy of globally-integrated box models described by Harvey (1989a). It is a 4-box model, with two living biomass boxes (where living biomass comprises woody material, leaves/needles, grass, roots, etc.) and two dead biomass boxes. One of the living biomass boxes is a rapid turnover box, assumed to have a zero time constant (left box in Fig. 10). A fraction of gross primary productivity (G) cycles through this box directly back into the atmosphere and may be ignored on time scales of interest here. The remaining part of the gross primary productivity is partitioned through the rapid turnover box into fractions which enter the other living biomass box (mass of box P, flux into box G_P), a detritus box (mass H, flux G_H) and a soil box (mass S, flux G_S). The mass-P box (or plant box) has G_P as its sole source term; its sinks are respiration (R), litter production (L: L_H to the detritus box and L_S to the soil box) and a land-use-change component (D_P). The detritus box has sources from gross primary productivity (G_H) and litter production (L_H); its sinks are fluxes to the atmosphere (Q_A), the soil box (Q_S) and land-use

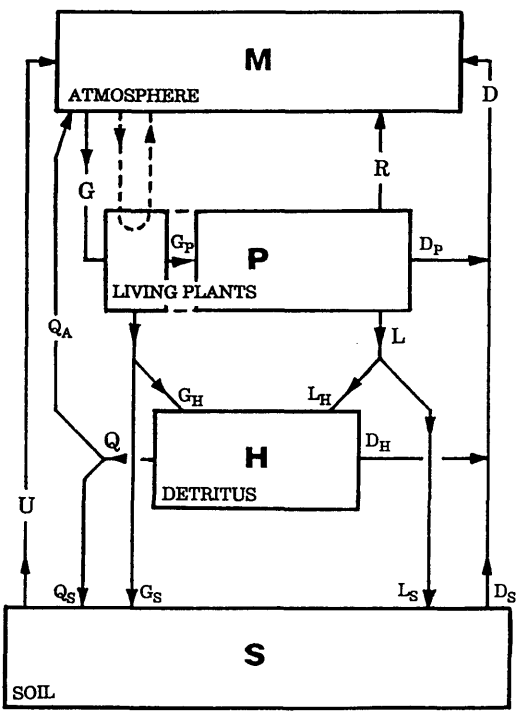

Fig. 10. The terrestrial component of the carbon cycle model.

changes (D_H). The soil box has sources from the detritus box (Q_S), litter production (L_S) and gross primary productivity (G_S); its sinks are through oxidation to the atmosphere (U) and land-use changes (D_S).

The equations describing mass balances in each box and in the atmosphere (mass M) are (cf. eqs. (2) and (3)):

$$\mathrm{d}M/\mathrm{d}t = I - F - \mathrm{d}B/\mathrm{d}t$$

$$= I - F - (G - R - Q_A - U - D_g), \tag{B1}$$

where $D_g = D_P + D_H + D_S$, F is the flux into the oceans (i.e., $\mathrm{d}O/\mathrm{d}t$ where O is the oceanic carbon mass) and I is the input from industrial emissions,

$$\mathrm{d}P/\mathrm{d}t = G_P - R - L - D_P, \tag{B2}$$

$$\mathrm{d}H/\mathrm{d}t = G_H + L_H - Q - D_H, \tag{B3}$$

where $Q = Q_A + Q_S$, and

$$\mathrm{d}S/\mathrm{d}t = G_S + Q_S + L_S - U - D_S. \tag{B4}$$

Note that D_g is gross deforestation, related to net deforestation by

$$D_n = D_g - \text{regrowth} = D_g - W. \tag{B5}$$

The sinks L, Q and U are assumed to be proportional to the P, H and S box masses respectively, via appropriate time constants:

$$L = P/\tau_P, \tag{B6}$$

$$Q = H/\tau_H, \tag{B7}$$

$$U = S/\tau_S. \tag{B8}$$

For the plant box, the term $L = P/\tau_P$ ensures that the total plant mass will relax back to its initial state if perturbed by a pulse land-use change (in the absence of changes in G_P and R). This relaxation acts as an effective regrowth term, so that D_P can only be interpreted as a component of gross land-use emissions. Gross and net land-use-change emissions are related by regrowth, W. Following Enting and Lassey (1993), this is determined by

$$W(t) = 1/\tau_P \int_0^t (D_P(u) - W(u))\,\mathrm{d}u,$$

which may be expressed differentially as

$$\mathrm{d}W/\mathrm{d}t = (D_P - W)/\tau_P. \tag{B9}$$

The fertilization effect is incorporated using the rectangular hyperbola expression for changes in NPP, as described in Appendix A. While this strictly applies only to NPP, the model requires separation of the effects on GPP and respiration. For simplicity it is assumed that

$$N/N_0 = G/G_0 = R/R_0, \tag{B10}$$

where N/N_0 is given in Appendix A.

The model parameters are now as follows: initial values are required for the plant, detritus and soil reservoirs (P_0, L_0 and S_0) and for GPP and respiration (G_0 and R_0); and partitioning factors are required for the flux out of the plant box (ϕP is assumed to go to the detritus, $(1-\phi)P$ to the soil), the flux out of the detritus (λL is assumed to go to the soil, $(1-\lambda)L$ to the atmosphere), the division of GPP ($g_P G$ to the plant box, $g_L G$ to the litter and $(1 - g_P - g_L)G$ to the soil), and the division of land-use-change emissions ($D_g = D_H + D_L + D_S = \gamma_P D_g + \gamma_L D_g + (1 - \gamma_P - \gamma_L) D_g$). Since the time step used for the model is one year, $g_L G$ represents the annual accumulation of GPP in the litter due to dying above-ground plant matter, while $(1 - g_P - g_L)G$ is the annual accumulation of GPP in the soil due to root turnover. Based on estimates in the literature (see, e.g., Harvey, 1989a) the following parameter values are used: $P_0 = 750$ GtC, $L_0 = 80$ GtC, $S_0 = 1450$ GtC, $G_0 = 76$ GtC/yr, $R_0 = 14$ GtC/yr, $\phi = 0.98$, $\lambda = 0.05$, $g_P = 0.35$, $g_L = 0.60$, $\gamma_P = 0.70$, $\gamma_L = 0.05$. Results are relatively insensitive to the precise choice of parameter values.

The time constants, τ_P, τ_L, and τ_S are determined fully by the other parameters and may be obtained directly from the initial (steady-state) forms of eqs. (B2), (B3), and (B4). The above parameter values lead to $\tau_P \approx 60$ yr, $\tau_H \approx 1.4$ yr and $\tau_S \approx 209$ yr.

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