Climate and interannual variability of the atmosphere-biosphere $^{13}$CO$_2$ flux

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[1] We present a bottom-up approach to simulate the terrestrial isotopic carbon variations using the Lund-Potsdam-Jena dynamic global vegetation model (LPJ-DGVM). LPJ is extended to include isotopic fractionation of $^{13}$C at the leaf level during assimilation and includes a full isotopic terrestrial carbon cycle. The model thus allows a quantitative analysis of the net biosphere exchange of CO$_2$ and $^{13}$CO$_2$ with the atmosphere as a function of changes in climate, atmospheric CO$_2$, and the isotope ratio of CO$_2$. LPJ simulates a global mean isotopic fractionation of 17.7‰ at the leaf level with interannual variations of ca. 0.3‰. Intertannual variability in the net $^{13}$CO$_2$ flux between atmosphere and terrestrial biosphere is of the order of 15 PgC‰ yr$^{-1}$. It is reduced to 4 PgC‰ yr$^{-1}$ if the leaf-level fractionation factor is held constant at the long term mean. Taking climate driven variable fractionation effects into account in double deconvolution studies we estimate that this could imply shifts of up to 0.8 PgC yr$^{-1}$ in the inferred partitioning between terrestrial and oceanic carbon sinks.


1. Introduction

[2] Over the last 50 years anthropogenic emissions of CO$_2$ added a total of ca. 210 petagrams (10$^{15}$ g) of carbon (PgC) into the atmosphere [Marland et al., 2001]. During the same period the atmospheric CO$_2$ content increased by 55 ppmv from 310 ppmv CO$_2$ in 1950 to 365 ppmv CO$_2$ in 1998 which is equivalent to ca. 115 PgC only [Prentice, 2001]. The remaining carbon has been absorbed by the oceans and the land biosphere. Changes in the concentration and $^{13}$C/$^{12}$C ratio of atmospheric CO$_2$ can be used to constrain the global carbon budget and derive location and magnitude of carbon sources and sinks [e.g., Keeling et al., 1989; Ciais et al., 1995; Heimann and Maier-Reimer, 1996]. A common method to infer the ocean and land contributions to these variations is the double-deconvolution of measured records of atmospheric CO$_2$ and $^8$δC ($^8$δC is calculated as the deviation with respect to a standard: $^8$δC = ($^13$C/$^12$C$_{atm}$ – $^13$C/$^12$C$_{std}$) / ($^13$C/$^12$C$_{std}$) × 1000‰) [Keeling et al., 1989; Francey et al., 1995; Joos and Bruno, 1998]. This “top-down” partitioning calculation is sensitive to various quantities, including the assumed isotopic signature of terrestrial carbon, which is usually held constant [e.g., Keeling et al., 1989; Joos and Bruno, 1998].

[3] However, the isotopic signature of the terrestrial biosphere is highly variable in space and time. The $^8$δC of assimilated CO$_2$ itself depends strongly on interannual climatic variability and on vegetation composition [Kaplan et al., 2002]. Because of their different anatomy, C$_4$-plants discriminate against heavy isotopes to a much smaller degree than C$_3$-plants. The double-deconvolution inversion method requires an a priori representation of the global $^{13}$CO$_2$ flux field. Our ability to characterize this $^{13}$CO$_2$ flux field in space and time based on field-scale measurements has not been possible. To date, most inversions have made simplifying assumptions to describe this flux field, e.g., a constant fractionation factor in space in time. Recently, several process-based model representations of the global $^{13}$CO$_2$ flux field have appeared [Fung et al., 1997; Kaplan et al., 2002]. However, none of these studies have demonstrated how interannual variability in climate affects plant physiology and vegetation dynamics, and through these processes, the $^{13}$CO$_2$ flux field and its, potentially large, implications for inverse modeling studies.

[4] Here, the dynamic global vegetation model LPJ is used to simulate CO$_2$ and $^{13}$CO$_2$ exchange between the atmosphere and biosphere for the last century in a consistent terrestrial carbon cycling framework. A set of experiments are performed to investigate the impact of the processes on the interannual variations in terrestrial carbon isotope discrimination, and to estimate the importance of considering a temporally-variable flux field for inverse modelling studies.

2. Methodology

2.1. The LPJ Dynamic Global Vegetation Model

[5] LPJ [Sitch et al., 2000] combines mechanistic treatment of terrestrial ecosystem structure (vegetation composition, biomass) and function (energy absorption, carbon cycling). Vegetation dynamics are updated annually based on the productivity, disturbance, mortality, and establishment of nine different plant functional types (PFTs). Modelled potential vegetation cover (including C$_3$-/C$_4$-plant distribution) depends on competition and climate history.
[6] The enhanced version of LPJ used here fully incorporates $^{13}$C as a tracer in the carbon cycle simulated by the model. The carbon isotope module is based on Kaplan et al. [2002] which has been validated with measurements at the leaf, canopy, and background atmosphere scales. LPJ, as the successor of the BIOME model used by Kaplan et al. [2002], is based on an almost identical formulation of ecosystem structure and function except for the vegetation dynamical elements. The module includes the calculation of isotope fractionation during photosynthesis, the addition of an extra above ground litter pool to represent the age spectrum of decomposing organic matter in a more realistic way, and separate accounting of all internal carbon pools for $^{12}$C and $^{13}$C.

[7] The isotope discrimination during assimilation is calculated as described by Lloyd and Farquhar [1994]. LPJ explicitly simulates the actual inter-cellular-to-atmospheric CO$_2$ concentration ($c_{cis}/c_a$) ratio through a coupled photosynthesis and water-balance canopy conductance scheme. Photosynthesis and carbon isotope discrimination are calculated on a daily time step representing daily average values; assimilated CO$_2$ and $^{13}$CO$_2$ is allocated to the four different tissue pools (leaves, sap- and heart-wood, roots) on an annual basis. Soil and litter total C and $^{13}$C pools are updated monthly. As isotope fractionation processes during respiration are poorly understood, no fractionation is assigned for the decomposition.

2.2. Atmospheric CO$_2$ and $^{13}$CO$_2$ Budget

[8] The atmospheric CO$_2$ budget is given by [e.g., Tans et al., 1993]

$$\frac{d F_a}{d t} = F_T + F_{ac, na} + F_{bu} = F_a + (F_{ba} - F_{ab})$$

(1)

where $F_a$ is the sum of net ocean and fossil fuel fluxes, $F_{ab}$ is the flux from the atmosphere to the biosphere (here taken as the net primary production, NPP), and $F_{ba}$ is the flux from the biosphere to the atmosphere, i.e., the heterotrophic respiration which here is the sum of the carbon emissions to the atmosphere from the litter and soil pools respectively, and the CO$_2$ flux due to fire disturbance. Correspondingly the atmospheric $^{13}$CO$_2$ budget can be formulated as

$$\frac{d^{13}F_a}{d t} =^{13}F_T + ^{13}F_{ac, na} + ^{13}F_{bu}$$

$$= ^{13}F_a + (F_{ba}R_b - F_{ab}\alpha_{ab}R_a)$$

(2)

where $R_a$ and $R_b$ are the isotopic ratios of atmospheric and biospheric carbon, respectively, and $\alpha_{ab}$ is the fractionation factor for the atmosphere-biosphere flux. Here, we assume that gross assimilation and autotrophic respiration are in isotopic equilibrium on an interannual time scale [Ekbollad and Högberg, 2001]. Focusing only on the time dependency of the isotopic ratio $R_a$ and considering only biospheric fluxes it follows with $F_{bu} = F_{ba} - F_{ab}$ that

$$\frac{d^{13}F_a}{d t} =^{13}F_T + ^{13}F_{ac, na} + ^{13}F_{bu}$$

$$= ^{13}F_a + (F_{ba}R_b - F_{ab}\alpha_{ab}R_a)$$

(3)

[9] Reformatting equation 3 in terms of the $\delta$-Notation (for nomenclature see e.g., Mook [1986]) and inserting a temporal mean (1950–98) fractionation term $\tau_{ab}$, the temporal change in $\delta^{13}$C from terrestrial biospheric fluxes can be described with three different terms:

$$\left[ N_a \frac{d^{13}C_a}{dt} \right]_{\text{biom}} = F_{bu}(\delta^{13}C_b - \delta^{13}C_a)$$

$$+ F_{ab}(\delta^{13}C_b - (\delta^{13}C_a + \tau_{ab})) + F_{ab}(\tau_{ab} - \epsilon_{ab})$$

(4)

The first term in equation 4 approximates the influence of the variability in the net terrestrial flux on the temporal change in $\delta^{13}$C, the second term represents the terrestrial $^{13}$C Suess-Effect and the third term $F_{ab}(\tau_{ab} - \epsilon_{ab}) = \Delta^{13}F_{ab}$ shows the influence of a time-varying fractionation factor on the global $^{13}$CO$_2$ budget. Global values are obtained by the flux-weighted sum of the quantities at each grid cell, which are then averaged over the globe.

2.3. Set-Up of Model Experiments

[10] Two experiments were performed at 0.5° spatial resolution. For both experiments (ISOVAR and ISOFIX) climate forcing data (monthly mean fields of temperature, precipitation, and cloud cover) was taken from the CRU/05 1901–1998 monthly climate time-series [New et al., 2000]. In the ISOVAR experiment discrimination was calculated according to the Lloyd and Farquhar [1994] scheme as described above, while for the ISOFIX experiment we prescribed a constant discrimination of 18.7‰ for C$_3$-plants and 3.4‰ for C$_4$-plants (global mean values from the ISOVAR experiment). For both experiments, we used the time-series of atmospheric CO$_2$ as described in McGuire et al. [2001] and an extended version of the atmospheric $\delta^{13}$C time-series of Francey et al. [1999].

3. Results and Discussion

[11] The modelled leaf discrimination varies from 3‰ to 4‰ for C$_4$-grasses and 10‰ to 23‰ for C$_3$-plants. This high variability of up to 13‰ in the leaf discrimination in C$_3$-PFTs reflects the sensitivity of the discrimination to climate conditions in different regions. The water stress induced by decreasing soil moisture and high water vapour pressure deficit controls the stomatal closure and therefore the $c/c_a$ ratio, which is the main variable governing the isotope discrimination. Figure 1 displays the spatial hetero-
Experiments fractionation experiment with constant climate but variable CO2, the climate had a direct effect on discrimination. In a third and the discrimination factor, or second a long-term trend in Tans et al. to 18.6
The value from the ISOVAR experiment for the years 1950 assimilated by C4-plants, this is partly because of the an amount of 25%, in LPJ less then 10% of total carbon is and Farquhar [1994] (reported value of 14.8
nation against 13C during photosynthesis, solid line
Global annual time-series of modelled discrimi-
SCHOLZE ET AL.: ATMOSPHERE-BIOSPHERE 13CO2 FLUX

Table 1. Mean Discrimination for Two Averaging Periods (1901–1930 and 1950–1998) [%], the Max. Interannual Variability in Discrimination [%], and in Δ13Fab [PgC %o yr⁻¹] for Both Experiments

<table>
<thead>
<tr>
<th>Exp.</th>
<th>Specifications</th>
<th>εab</th>
<th>IAV εab</th>
<th>IAV Δ13Fab</th>
</tr>
</thead>
<tbody>
<tr>
<td>ISOVAR</td>
<td>var. frac.</td>
<td>17.40</td>
<td>17.67</td>
<td>0.3</td>
</tr>
<tr>
<td>ISOFIX</td>
<td>const. frac.</td>
<td>17.72</td>
<td>17.75</td>
<td>0.1</td>
</tr>
</tbody>
</table>

generality of the global pattern in the modelled leaf discrimination. The impact of arid climates on discrimination is clearly visible in regions such as central North America and southern South America, Central Asia and in parts also Australia. The influence of C₄-photosynthesis can mainly be seen in the eastern Brazilian grasslands, the African sub-
tropics and the northern region of Australia with discrimination values between 10%o and 12%o.

[12] The mean globally averaged flux weighted isotopic fractionation εab over the years 1901–1930 and 1950–1998, the maximum interannual variability of εab, and of Δ13Fab are summarized in Table 1 for both experiments. The value from the ISOVAR experiment for the years 1950 to 1998 of 17.7%o is within the range of values from 14.8%o to 18.6%o reported in recent studies [Keeling et al., 1989; Tans et al., 1993; Lloyd and Farquhar, 1994; Fung et al., 1997; Kaplan et al., 2002]. Differences arise especially from different amounts of C₄-photosynthesis: while Lloyd and Farquhar [1994] (reported value of 14.8%o) prescribe an amount of 25%, in LPJ less then 10% of total carbon is assimilated by C₄-plants, this is partly because of the absence of C₄ land use (pastures and crops). In the ISOFIX experiment, the mean fractionation factor for the period 1950 to 1998 is 17.8%o, almost identical to the ISOVAR experiment.

[13] The interannual variability of εab in ISOFIX is greatly reduced to 0.1%o compared to 0.3%o in ISOVAR as a consequence of the constant fractionation factor (see Figure 2). This remaining variability of ε≈0.1%o is entirely caused by changes from C₃- to C₄-plants and their productiv-
ty. The long-term trend in discrimination in the ISOVAR experiment could have two possible explanations: first increasing atmospheric CO₂ content changed the ε/εᵣ ratio and the discrimination factor, or second a long-term trend in climate had a direct effect on discrimination. In a third experiment with constant climate but variable CO₂, the discrimination factor did not show any trend over the 98 years. Therefore we assume that the trend in the ISOVAR experiment is mainly a climate effect and most likely a response of the plants to increased water stress due to global warming during the 20th century. For the same time period, McGuire et al. [2001] already found a small declining trend in NPP leading to net carbon loss when considering climate only.

[14] Globally averaged time-series of the three terms yielding the temporal variations in atmospheric δ13Cf from equation 4 and their corresponding CO₂ fluxes (Fab and Fbnet, top panel) are plotted in Figure 3. The net atmos-
phère-biosphere flux (Fbnet) shows mainly an uptake of carbon by the land biosphere for the last 50 years, the amount of uptake is comparable to other model studies [McGuire et al., 2001]. Interannual variability in the net flux is related to the variability in the El Niño Southern Oscillation (ENSO) [Keeling et al., 1989]. As Fbnet is the same for both experiments, the values from the first term representing the influence of the net flux (upper middle panel) are almost identical for both experiments.

[15] A secular increasing trend in the terrestrial Suess-Effect from 17.5 PgC %o yr⁻¹ in 1950 to more than 30 PgC %o yr⁻¹ in 1998 caused by the invasion of isotopically light anthropogenic CO₂ into the biosphere is simulated (lower middle panel in Figure 3). High frequency (interannual) variability is strongly reduced as the 13C entering the terrestrial biosphere is damped in the δ13Cₑ emitted from

Figure 2. Global annual time-series of modelled discrimination against 13C during photosynthesis, solid line ISOVAR and dashed line ISOFIX.

Figure 3. Global annual time-series of modelled Fbnet (solid line, negative values indicate net storage in land biosphere) and Fab (dashed line), top panel, and the three terms giving the time dependency of δ13Cₑ (see equation 4; in the following: solid ISOVAR and dashed ISOFIX): upper middle panel influence of the net flux (first term), lower middle panel Suess-Effect (second term) and bottom panel Δ13Fab (third term).
the biosphere because of the relatively long residence time of carbon in the land biosphere. The offset between the two experiments is caused by the difference in the fractionation factor at the beginning of the simulation period as mentioned above.

[16] The bottom panel of Figure 3 displays $\Delta^{13}F_{ab}$. Calculating $\Delta^{13}F_{ab}$ using the mean global average $\varepsilon_{ab}$ for the years 1920–1995 gives the magnitude of changes in the $^{13}$CO$_2$ fluxes due to climate variability in the fractionation factor $\varepsilon_{ab}$ opposed to a constant $\varepsilon_{ab}$. In Table 1 the differences in the interannual variability of $\Delta^{13}F_{ab}$ are summarized. In the ISOFIX experiment the variations are only about 4 PgC $\%_\circ$ yr$^{-1}$, whereas the ISOVAR experiment with climatic dependent discrimination exhibits interannual variations up to 15 PgC $\%_\circ$ yr$^{-1}$. The difference in the variability in $\Delta^{13}F_{ab}$ between the two experiments is controlled only by the impact of climate variability on isotope discrimination during photosynthesis. In general, interannual variability is especially pronounced during pronounced ENSO events (e.g., 1992/93 and 1997/98) which lead through widespread drought conditions to a decrease in the isotopic fractionation factor.

[17] Using a variable climatic dependent discrimination in double deconvolution studies would certainly lead to a modification of the derived terrestrial sinks. Taking a variable fractionation factor into account we estimate that this could lead to a year-to-year shift of up to 0.8 PgC yr$^{-1}$ in the inferred partitioning between terrestrial and oceanic carbon sinks. In general, a transient decrease in the terrestrial isotopic discrimination factor would imply a corresponding reduction in the inferred terrestrial source by means of a double deconvolution calculation.

4. Conclusions

[18] This study demonstrates that large interannual variations in $^{13}$C fractionation, induced by anomalous climate events (e.g., ENSO) have a strong impact on the globally integrated terrestrial $^{13}$CO$_2$ fluxes. These fluctuations result primarily from plant physiological responses and to a minor extent from changes in vegetation composition (shifts between C$_3$- and C$_4$-plants) induced by climate variations. Ignoring these effects in global double deconvolution studies may result in significant errors in the inferred annual net atmosphere-biosphere flux of up to 0.8 PgC/yr, more than half of the global estimated biospheric sink during the 1990’s [Francey et al., 1995]. So far, these simulations do not consider land use change which influences the global discrimination values mainly through tropical C$_4$-pastures [Townsend et al., 2002] and C$_4$-crops, e.g., maize cultivation in the northern hemisphere. However, we assume that the effect of land use change on the interannual variability is small compared to the climate effect. Future inverse modeling studies, which may lead to important revision of published carbon budgets for the past two decades, may wish to use the isotope LPJ framework for prescribing variations in terrestrial isotopic discrimination.

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References


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