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Interannual variability of the global carbon cycle (1992–2005) inferred by inversion of atmospheric CO₂ and $\delta^{13}\text{CO}_2$ measurements

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[1] We present estimates of the surface sources and sinks of CO₂ for 1992–2005 deduced from atmospheric inversions. We use atmospheric CO₂ records from 67 sites and 10 $\delta^{13}\text{CO}_2$ records. We use two atmospheric models to increase the robustness of the results. The results suggest that interannual variability is dominated by the tropical land. Statistically significant variability in the tropical Pacific supports recent ocean modeling studies in that region. The northern land also shows significant variability. In particular, there is a large positive anomaly in 2003 in north Asia, which we associate with anomalous biomass burning. Results using $\delta^{13}\text{CO}_2$ and CO₂ are statistically consistent with those using only CO₂, suggesting that it is valid to use both types of data together. An objective analysis of residuals suggests that our treatment of uncertainties in CO₂ is conservative, while those for $\delta^{13}\text{CO}_2$ are optimistic, highlighting problems in our simple isotope model. Finally, $\delta^{13}\text{CO}_2$ measurements offer a good constraint to nearby land regions, suggesting an ongoing value in these measurements for studies of interannual variability.

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1. Introduction

[2] This paper presents an update of the work of Rayner *et al.* [1999]. It reflects several improvements in both inversion methodology and data density. It also covers a more recent period than that study. The most important improvement is the refinement and assessment of calibration scale propagation over decadal time frames for the CSIRO CO₂ isotope records [Allison and Francey, 2007]. This provides a robust foundation for CSIRO $\delta^{13}\text{CO}_2$ records with reasonable global coverage over the whole period. The ability of $\delta^{13}\text{CO}_2$ data to separate certain terrestrial fluxes from other CO₂ fluxes has made it a common tool in atmospheric inversions [e.g., Tans *et al.*, 1993; Ciais *et al.*, 1995; Enting *et al.*, 1995] but its use requires care. The other methodological improvements are described in section 2.

[3] The study period encompasses two of the most dramatic events seen in the nearly five decades of atmospheric CO₂ concentration measurements: the near-flattening of the atmospheric CO₂ growth rate centered on 1993 and the large spike in this growth rate in 1997–1998. These have been

treated in previous inversion papers [e.g., Rödenbeck *et al.*, 2003a; Peylin *et al.*, 2005a; Baker *et al.*, 2006]. None of these studies used $\delta^{13}\text{CO}_2$ data so we can test whether this data set challenges previous conclusions. Our study period also includes the recent sustained high growth rates seen in 2002–2003. There is currently little published information on the spatial structure associated with these anomalies although Van der Werf *et al.* [2006] describe anomalies in some key processes.

[4] The outline of the paper is as follows. In section 2 we discuss the changes made to the methodology from Rayner *et al.* [1999]. Section 3 presents the major results, concentrating on interannual variability. Section 4 carefully evaluates the role played by the $\delta^{13}\text{CO}_2$ measurements in the inversion as well as the various assumptions we make to use them.

2. Method

[5] We use the same basic method, Bayesian Synthesis Inversion, as Rayner *et al.* [1999] (see Enting [2002] for a detailed explanation). We construct the cost function

$$\chi^2 = \frac{1}{2} \left[(\vec{s} - \vec{s}_0)^T \mathbf{C}(\vec{s}_0)^{-1} (\vec{s} - \vec{s}_0) + (\mathbf{J}\vec{s} - \vec{d})^T \mathbf{C}(\vec{d})^{-1} (\mathbf{J}\vec{s} - \vec{d}) \right], \quad (1)$$

where $\vec{s}$ represents sources, $\vec{s}_0$ prior source estimates, $\vec{d}$ the observed concentrations and isotopic values and $\mathbf{J}$ is the Jacobian matrix of sensitivities of observations with respect

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to $\vec{s}$. Uncertainties on $\vec{d}$ and $\vec{s}_0$ are expressed in covariance matrices $\mathbf{C}$. $\mathbf{J}$ is calculated by repeated runs of a transport model, one for each region and month. The minimization is performed with the singular value decomposition [Enting, 2002, p. 66]. We also compute the posterior covariance of fluxes following Enting [2002, section 10.3]. Part of the motivation for this work is an exploration of the impact of $\delta^{13}\text{CO}_2$ measurements on estimates of interannual variability. It is convenient, therefore, to choose a setup comparable with the TransCom study of interannual variability performed by Baker *et al.* [2006] (denoted T3-LAV). Thus many of the details for the setup are chosen to make it compatible with that of Baker *et al.* [2006].

2.1. Transport Models

[6] We have performed the inversion using two different transport models, since it is valuable to have some measure of the sensitivity of the estimated fluxes to atmospheric transport. The first model is a version of the Model of Atmospheric Transport and Chemistry (MATCH) [Rasch *et al.*, 1997] driven by winds from the Middle Atmosphere Community Climate Model version 2 (MACCM2) [Boville, 1995]. The model was modified at the Cooperative Research Centre for Southern Hemisphere Meteorology and this version is designated CRC-MATCH. The resolution is 5.6° longitude by 2.8° latitude by 24 levels (some MACCM2 stratospheric levels were not used). The model is described in detail by Law and Rayner [1999] and participated in annual mean and interannual TransCom inversions [Gurney *et al.*, 2002, 2004; Baker *et al.*, 2006] labeled MATCH-MACCM2.

[7] CRC-MATCH exhibited behavior intermediate among the models in TransCom. It showed an intermediate large-scale concentration gradient arising from the fossil fuel source and a high strength of the covariance between seasonal transport and seasonal terrestrial sources, the so-called rectifier effect [Keeling *et al.*, 1989; Denning *et al.*, 1995]. Its inverted fluxes were also near the center of the TransCom range for both annual mean and seasonal cycle cases.

[8] The second model used is the CSIRO Conformal-Cubic Atmospheric Model (CCAM). It is an atmospheric general circulation model [McGregor, 1996; McGregor and Dix, 2001] with tracer transport by advection and convection occurring online. Here the model is run with approximately uniform resolution globally with a model grid spacing of around 220 km. The model can run independently generating its own climate or it can be nudged to forcing fields. Here we nudge to NCEP [Kalnay *et al.*, 1996; Collier, 2004] 6 hourly horizontal wind fields (u and v) for 1999–2000. The model is described in more detail by Law *et al.* [2006] and participated in the annual mean TransCom inversion [Gurney *et al.*, 2003], labeled as CSIRO. It showed intermediate behavior for both the north-south gradient due to fossil fuel emissions and for the biosphere rectifier.

2.2. Source Resolution

[9] Rayner *et al.* [1999] used a relatively coarse description of sources with 14 land and 12 ocean regions while Baker *et al.* [2006] used 11 land and 11 ocean regions. Here

we expand the source description to 67 land and 49 ocean regions for CRC-MATCH and 94 land and 52 ocean regions for CCAM. Kaminski *et al.* [2001] found these resolutions sufficient to avoid aggregation error. We also add presubtracted fields for the seasonal biosphere [Randerson *et al.*, 1997] and ocean fluxes [Takahashi *et al.*, 1999]. We use two patterns for fossil fuel combustion [Andres *et al.*, 1996; Brenkert, 1998] (available from <http://cdiac.esd.ornl.gov/ndps/ndp058a.html>) corresponding to the years 1990 and 1995 with annually varying magnitudes following Baker *et al.* [2006, Table 1]. After 2003 we used a 3% annual extrapolation of fossil magnitude.

[10] Prior estimates for regional ocean fluxes, which are corrections to the background flux, are all zero. We have distributed the land use change flux used by Gurney *et al.* [2002] uniformly in time and spatially according to the CASA estimate of net primary productivity (NPP) [Randerson *et al.*, 1997]. The prior uncertainties are informed by Baker *et al.* [2006]. For CRC-MATCH we have chosen a discretization of regions that maps directly onto the 22 regions of that study and chosen prior uncertainties so that the total uncertainties on those regions are equal to those of Baker *et al.* [2006] and are consequently seasonally invariant for ocean but variable for land. Within these 22 regions we allocate uncertainty according to CASA NPP for land and area for ocean. Ocean uncertainties on monthly fluxes, range from $4 \text{ gC m}^{-2} \text{ yr}^{-1}$ for a region in the Mediterranean to $80 \text{ gC m}^{-2} \text{ yr}^{-1}$ for a region in the south Atlantic with an RMS of $47 \text{ gC m}^{-2} \text{ yr}^{-1}$. Over land, uncertainties range from $21 \text{ gC m}^{-2} \text{ yr}^{-1}$ over Greenland to $2400 \text{ gC m}^{-2} \text{ yr}^{-1}$ over New Zealand with an RMS of $517 \text{ gC m}^{-2} \text{ yr}^{-1}$. We impose no correlations among flux components. For CCAM the source regions did not coincide exactly with the TransCom boundaries but the uncertainties were chosen to be similar to those used for the CRC-MATCH case.

2.3. Treatment of $^{13}\text{CO}_2$

[11] The strong fractionation of C_3 photosynthesis against $^{13}\text{CO}_2$ means we can use measurements of $\delta^{13}\text{CO}_2$ to separate net fluxes of C_3 plant material from those of C_4 plants or oceans [e.g., Tans *et al.*, 1993; Enting *et al.*, 1995; Rayner *et al.*, 1999]. There are two unrelated parts to the inclusion of $\delta^{13}\text{CO}_2$ in an atmospheric inverse model. First, what components do we use to describe the $^{13}\text{CO}_2$ fluxes and how do we relate them to net carbon fluxes? Secondly, how do we describe the evolution of $\delta^{13}\text{CO}_2$ in the atmosphere? We will deal with each of these in turn.

2.3.1. Modeling $^{13}\text{CO}_2$ Source Components

[12] Following Tans *et al.* [1993], we model the flux of $^{13}\text{CO}_2$ in region i and time t as a sum of gross fluxes that affect only $^{13}\text{CO}_2$ and net fluxes which are linked to CO_2 fluxes:

$$f^{\text{C}^{13}}(i, t) = g^{\text{C}^{13}}(i, m(t)) + (y(t) - y_0)h^{\text{C}^{13}}(i, m(t)) + \alpha^{\text{C}^{13}}(i)f^{\text{CO}_2}(i, t), \quad (2)$$

where $m(t)$ is the month of year corresponding to time t , $y(t)$ the year corresponding to time t and y_0 the first year of the inversion. $g^{\text{C}^{13}}$, $h^{\text{C}^{13}}$ and f^{CO_2} are unknowns in the

optimization and $\alpha^{C^{13}}$ is a fractionation factor in permil. Thus we model the gross fluxes as a seasonally varying component ($g^{C^{13}}$) plus a seasonally varying and linearly increasing term ($h^{C^{13}}$) to account for possible changes in disequilibrium. For monthly resolution of fluxes this treatment introduces 24 extra variables for each region into the description of $^{13}\text{CO}_2$ fluxes. We give $g^{C^{13}}$ and $h^{C^{13}}$ large prior uncertainties. The formulation means $g^{C^{13}}$ and $h^{C^{13}}$ will absorb all information from $\delta^{13}\text{CO}_2$ data corresponding to constant or linearly evolving $^{13}\text{CO}_2$ sources. Thus $\delta^{13}\text{CO}_2$ data will not inform long-term mean CO_2 sources.

[13] $\alpha^{C^{13}}$ represents the spatial variation in isotopic fractionation of the net CO_2 flux. Its value is based on the proportions of C_3 and C_4 plants within a given region according to the expression

$$\alpha_i^{C^{13}} = -17.5 + 16 \times F_i, \quad (3)$$

where F_i is the fraction of C_4 vegetation in a region. F_i is computed from the distribution of *Still et al.* [2003a].

[14] Any process that affects $^{13}\text{CO}_2$ fluxes without affecting CO_2 fluxes but which cannot be modelled by our linear model will be misattributed in the inversion. Such processes include changes in leaf-level fractionation or in relative productivity of C_3/C_4 ecosystems [e.g., *Randerson et al.*, 2002; *Scholze et al.*, 2003; *Still et al.*, 2003b] or in the proportion of C_3 and C_4 vegetation involved in biomass burning [*Van der Werf et al.*, 2006]. *Scholze et al.* [2003] estimated an error of 0.5 PgCy^{-1} for inversions neglecting the fractionation effects. Presumably the error is larger when we consider the full ensemble of missing effects. This is a significant error at global scale but is likely to be smaller at regional scales since we do not expect errors to cancel spatially. We can compare this with uncertainty estimates returned by the inversion. Also, neglecting these effects should degrade the fit to $\delta^{13}\text{CO}_2$ data compared to that for CO_2 . In section 4 we will compare the ability of the inversion to fit CO_2 and $\delta^{13}\text{CO}_2$ data.

2.3.2. Modeling $^{13}\text{CO}_2$ in the Atmosphere

[15] There is an equilibration of an atmospheric $^{13}\text{CO}_2$ anomaly with the underlying reservoirs which is separate from the equilibration of a concomitant CO_2 anomaly. To see this, imagine the extreme case with two reservoirs with CO_2 concentrations at equilibrium but all the $^{13}\text{CO}_2$ in one reservoir. The imbalance in $^{13}\text{CO}_2$ between the reservoirs will be reequilibrated by gross exchange even though there is no net CO_2 flux. Thus in modeling the response of atmospheric $^{13}\text{CO}_2$ we must modify the transport Jacobians to account for these exchanges. The exchanges can be parameterized by inserting pulses of $^{13}\text{CO}_2$ into the atmosphere of a model of surface exchange and biogeochemistry. This has been calculated for various ocean models by *Joos et al.* [1996] and for one terrestrial model by *Thompson and Randerson* [1999]. We are interested in the total atmospheric response so we need a model combining ocean and terrestrial responses. We use a parameterization based on the simple global model of *Trudinger et al.* [1999]:

$$R_{C^{13}} = R_{\text{CO}_2} \left[0.7575e^{-t/2.236} + 0.2153e^{-t/22.18} \right], \quad (4)$$

where $R_{C^{13}}$ and R_{CO_2} are the response functions for pulses of $^{13}\text{CO}_2$ and CO_2 respectively, used to form the transport Jacobian, and t is the time in years.

2.4. Data

[16] We use a set of 67 CO_2 concentration records taken from *GLOBALVIEW-CO₂* [2006]. We use a subset of the stations used by *Baker et al.* [2006]. We delete all duplicate records at a site taking, in general, the longer record. The 67 sites are shown as the circles in Figure 1a. *Baker et al.* [2006] chose records with more than 68% coverage during their study period and increased the data uncertainty during periods using extrapolated data. We have a different study period than *Baker et al.* [2006] so do not always meet the 68% criterion. We use the same algorithm for computing data uncertainty as *Baker et al.* [2006]. Our CO_2 data uncertainties range between 0.3 ppm (the imposed minimum) and 9.4 ppm with 90% less than 2 ppm.

[17] We use ten records of atmospheric $\delta^{13}\text{CO}_2$ from the CSIRO Global Air Sampling Laboratory (GASLAB) [*Francey et al.*, 1996]. The calibration methods underpinning the data are described by *Allison and Francey* [2007]. We use the longest records from GASLAB as well as sites for which there is CO_2 data. We use the procedure of *Thoning et al.* [1989] to generate monthly means from irregular data using an 80-day cutoff. This is the same procedure, with the same filter parameters, used for the CO_2 data. We use the residual standard deviation (RSD) from the smoothed curve as the uncertainty on each monthly measurement with two exceptions. We impose a value of 0.05‰ if there is only one measurement in the month and we impose a minimum value of 0.015‰. The highest uncertainties are 0.05‰ with 90% less than 0.02‰. Unlike CO_2 we do not use extrapolated data for $\delta^{13}\text{CO}_2$ since we do not have enough records to define a marine boundary layer value.

2.5. Experiments

[18] We carried out two separate experiments with each of the two transport models. The control case uses data and uncertainties as described above while the second neglects $\delta^{13}\text{CO}_2$ data. We solved for flux magnitudes for each month in the period 1990–2005. Data was only introduced in 1992 so the first 2 years of fluxes were considered spin-up and are not used in subsequent analysis. For experiments using $\delta^{13}\text{CO}_2$ measurements we also included the $^{13}\text{CO}_2$ fluxes from equation (2).

3. Results

[19] We divide our treatment of results into their long-term mean and interannually varying components. We invest little space in the treatment of the long-term mean since our choices of data and the modeling of $\delta^{13}\text{CO}_2$ are optimized for inference of interannual variability. Also, *Baker et al.* [2006] have shown that, with the current state of atmospheric transport modeling, the interannual variability of fluxes is a more robust outcome of inversions than the long-term mean.

3.1. Long-Term Mean

[20] Figure 1 shows the mean CO_2 flux for 1992–2005 from the control inversions for CRC-MATCH and CCAM.

Figure 1. Maps of the mean posterior flux for 1992–2005 ($\text{gC m}^{-2} \text{yr}^{-1}$) excluding fossil fuel for the control cases from (a) the Cooperative Research Centre version of Model of Atmospheric Transport and Chemistry (CRC-MATCH) and (b) the Conformal-Cubic Atmospheric Model (CCAM). Locations of CO_2 data are shown in Figure 1a and $\delta^{13}\text{CO}_2$ data in Figure 1b. Note that the CRC-MATCH inversion does not solve for the Arctic or Antarctic fluxes.

The network of sites used in the inversion is also shown (CO_2 sites (Figure 1a) and $^{13}\text{CO}_2$ sites (Figure 1b)). The mean fluxes are broadly similar from the two inversions. Over land, both models suggest the tropics is generally a source with a sink at northern midlatitudes and a source in high latitudes. Flux uncertainties on land preclude a discussion of the long-term mean at much finer detail than this. The ocean fluxes from both inversions are similar, being dominated by the Takahashi background fluxes. The most noticeable difference is in the Pacific sector of the southern ocean where CCAM gives a source and CRC-MATCH a sink.

[21] We have compared the long-term mean from the CRC-MATCH inversions with the model mean from *Baker et al.* [2006] (T3-IAV). Recall that the CCAM inversion does not map onto the same regions as T3-IAV. The different study periods make comparison difficult. Here

we compare the period 1992–2005 from CRC-MATCH and 1992–2003 from T3-IAV. Generally the inversions agree for well-constrained regions, the largest exception being Europe where T3-IAV predicts a sink of 1.0 PgCy^{-1} while CRC-MATCH predicts a sink of 0.3 PgCy^{-1} . The difference is most likely due to the higher source resolution of the CRC-MATCH inversion which produces a mixture of sources and sinks impossible in the T3-IAV setup.

3.2. Flux Variability

[22] Figure 2 shows the 11-month running mean anomalies for the two control inversions and those of the T3-IAV model mean. Figures 2a–2f show fluxes for the northern extratropics, tropics and southern extratropics and for land and ocean separately. We define the boundaries between regions at 30° latitude. The fluxes plotted exclude all the

Figure 2. Eleven-month running mean flux anomaly for the CRC-MATCH (black, solid), CCAM (black, dashed), and *Baker et al.* [2006] T3-IAV (red) inversions for (a) northern land, (b) northern ocean, (c) tropical land, (d) tropical ocean, (e) southern land, and (f) southern ocean. The shaded area shows the $\pm\sigma$ uncertainty envelope from the CRC-MATCH inversion. The blue dotted line (Figures 2a–2d only) is a measure of the Southern Oscillation Index (SOI) scaled to fit the y -range of each panel. Note that the plotted range for tropical land (Figure 2c) is 3 times larger than the range in the other panels.

background fluxes so that the gradual trend in the fossil flux will not appear in these plots. The anomalies are constructed by subtracting the long-term mean for each month from the raw time series and constructing the 11-month running mean on the resultant time series. These two steps can be

combined into a single linear operator and we apply exactly the same operator to the posterior covariance matrix to obtain the $1\text{-}\sigma$ confidence intervals for the CRC-MATCH inversion shown as the shaded area. Finally, all three inversions use different regional patterns for the sources.

To ensure a proper comparison we project all estimates onto a spatial grid according to their own basis functions then calculate spatial integrals from this grid. This integration can exacerbate apparent differences among models if region boundaries coincide with large spatial gradients of sources.

[23] The large-scale features of the variability are somewhat familiar. In common with inversion studies since that of *Bousquet et al.* [2000], we obtain the greatest variability in the tropical land (note that Figure 2c is drawn to a larger scale). The variability for all three inversions is similar except for the southern land where T3-IAV is less variable.

[24] The southern land shows relatively small variability (consistent with the small land area south of 30°S) with the largest positive excursions in 1994 and 1997 and a negative anomaly in 2001. Variability in the austral ocean appears anticorrelated with that for southern land but is also generally small.

[25] The tropical land shows the well-known positive excursions in 1994 and 1997–1998 [e.g., *Bousquet et al.*, 2000; *Rödenbeck et al.*, 2003a] plus positive anomalies in late 2002 and 2005. We see negative anomalies spaced roughly evenly around these positive anomalies. The amplitudes of these excursions are larger than for other regions. There is generally good agreement between the models and with T3-IAV. The differences between the CRC-MATCH and CCAM fluxes in 1993 and 2001 mirror differences seen in the northern land region and are due to large flux gradients in Asia and the choice of region boundary.

[26] The tropical ocean shows smaller variability than the land. The two largest anomalies, a source from 1992–1994 and the large anomalous sink of 1997 are anticorrelated with tropical land anomalies. T3-IAV fluxes show an anomalous source in 2002–2003 but this is not shown by either control inversion.

[27] The northern land shows much structure. Before 1994 it is an anomalous sink. This is followed by moderate although short-lived excursions which are not always consistent between the models. Finally there is a large positive excursion from mid-2002 to mid-2003. The northern land anomaly in 2003 is the major contributor to the global anomaly in CO₂ growth rate previously noted by *Jones and Cox* [2005]. The northern ocean shows much less variability and is generally anticorrelated with the land. This is one region where the two control inversions show significantly different variability to T3-IAV.

[28] We assess the statistical significance of the variability by posing the null hypothesis “what is the chance that a series with zero variability but the given posterior uncertainty covariance would generate a realization with variability greater than that observed?” *Baker et al.* [2006] used the average value of the posterior uncertainty and a χ^2 test to calculate this. Here we use a Monte Carlo method based on realizations of a series with zero mean and temporal covariance given by the posterior covariance matrix spatially integrated and temporally smoothed as described above. The method accounts properly for the temporal uncertainty correlations introduced both by the inversion and the smoothing. The tropical land, northern ocean and northern land are all significant at the 95% level. If we decompose into ocean basins and continents, only the tropical Pacific

shows statistically significant variability under the same test.

[29] We test the significance of differences in variability with the same Monte Carlo method. The variability (measured by the standard deviation of smoothed anomalies) of the tropical land is about three times that for the tropical ocean. This could arise from the greater uncertainty of land versus ocean fluxes. We compared the variability of many realizations of flux with zero mean and uncertainty given by the posterior uncertainty covariance for tropical land and ocean. Less than 0.1% of the land realizations had more than three times the variability of their ocean counterparts, suggesting the difference in variability does not arise by chance. The ratio is, however, partly conditioned by the prior uncertainties. In a poorly constrained region like the tropics it is possible to trade-off variability between the land and ocean. One setting which can affect this trade-off is the relative prior uncertainty of land and ocean fluxes. Thus we can assert the statistical significance of the ratio of variability of land and ocean fluxes but not the robustness of the ratio.

[30] The statistical significance gives an interesting view of the anticorrelation noted above between land and ocean fluxes in the extratropics. Such anticorrelations have been noted at global scales by *Francey et al.* [1995] and *Keeling et al.* [1995] and spatially by many authors [e.g., *Rayner et al.*, 1999; *Bender et al.*, 2005; *Baker et al.*, 2006]. They have been ascribed variously to problems with $\delta^{13}\text{CO}_2$ records [*Francey et al.*, 1995], modeling of $^{13}\text{CO}_2$ [e.g., *Randerson et al.*, 2002], or to a lack of resolution in the inversion. The lack of resolution manifests itself as a relatively good constraint on the fluxes from the complete latitude band but an inability to separate the fluxes into land and ocean components. Following the work of *Allison and Francey* [2007], we are confident that the $\delta^{13}\text{CO}_2$ records show little spurious variability although, as we will see, we should be less confident in our ability to model them. Thus our anticorrelation is most likely a result of lack of resolution. Where anticorrelations occur in the absence of separate, significant variability, we should not spend much time seeking explanations for the variations.

[31] The pale blue dotted line in Figure 2 shows a measure of the Southern Oscillation Index (SOI) taken from <http://www.bom.gov.au/climate/current/soihtml1.shtml>. The SOI values are rescaled to fit the range of each plot and smoothed as for all other curves. The tropical land shows the expected response with SOI and flux changes in counterphase except for the early 1990s which we discuss below.

[32] Figure 3 shows the tropical Pacific anomalies, SOI and the model results of *Buitenhuis et al.* [2006] for the tropical Pacific, using their run where nutrients are not restored below the mixed layer. All curves are smoothed with an 11-month running mean. The relationship between SOI and flux in the tropical Pacific is smaller and more ambiguous than that for the land. The expected in-phase response holds between about 1995 and 2002, most clearly in the positive SOI events of 1996–1997 and 1998–1999 and the celebrated negative event of 1997–1998. Outside this we see a counterphase response to the negative event of

Figure 3. Eleven-month running mean flux anomaly for the tropical Pacific for the CRC-MATCH (black, solid), CCAM (black, dashed), and T3-IAV (red) inversions and for the ocean model fluxes of Buitenhuis *et al.* [2006] (green, dashed). The blue dotted line is a measure of the SOI scaled to fit the y-axis range.

the early 1990s and almost no response to the negative event of 2002–2003. After 1994, the inversion results and ocean model results are in good agreement, especially the small response in 2002. The lack of ocean response in 2002–2003 is striking and serves to amplify the impact of that ENSO event on the global growth rate. We also see little response in 2004 (in either ocean model or inversions) despite a moderate ENSO event.

[33] The northern extratropical fluxes, either for land or ocean, show considerable structure and statistically significant interannual variability. Northern ocean fluxes show a similar relationship to the SOI as tropical ocean fluxes while there is no clear relationship for the northern land. We also performed a simple correlation analysis of the northern flux anomalies and an index of the North Atlantic Oscillation from <http://www.cru.uea.ac.uk/cru/data/nao.htm>. We saw weak correlations with land and ocean fluxes over the latitude band although they were a little stronger with Europe and North America.

[34] We have noted in several regions the unusual behavior of the carbon cycle during the early 1990s. In our inversions it is characterized by large anomalous sinks in tropical and northern land and weaker offsetting sources in the oceans, particularly the tropical Pacific. The anomalous behavior is usually ascribed [e.g., Peylin *et al.*, 2005a] to the impact of the Mt. Pinatubo eruption. Peylin *et al.* [2005a] commented on the ambiguous attribution of the anomalous sink between the northern and tropical land. Here we see a strong response in the tropics, peaking in late 1992 and a delayed and weaker response in the northern extratropics. The tropical land response is surprising when we note the usual impact of the negative SOI event which would suggest a significant source. The results support the model study of Jones and Cox [2001] which attributed the dom-

inant response to the Pinatubo eruption to the tropical land. We cannot distinguish the various proposed mechanisms for the anomalous sinks but the 3-month running mean (not shown) indicates that the predominant anomalies in the northern extratropics occur in summer. The offsetting sources in the ocean are most likely from the lack of resolution in the inversion.

3.3. Northern Land Anomalies

[35] The requirement for considerable spatial and temporal smoothing of CO₂ fluxes (as we used in Figure 2) is a function of the weak constraint offered by the sparse observing network and consequent noise in retrieved fluxes. However data density is increasing so we attempt here a more detailed analysis of some large events in the best observed part of the globe and the temporal record, the northern hemisphere since 2001. We focus on the land regions because of their much larger variability. To allow better localization of events in time we smooth the flux anomalies with a triangular filter of full width 5 months. This filter also has superior properties regarding spectral aliasing to the running mean normally used.

[36] Figure 4 shows flux anomalies for 2001–2005 for three land regions north of 30°N, Europe, North Asia and North America. We divide Europe and North Asia at 60°E. Note that Europe includes parts of North Africa and the Middle East but the relatively low productivity of these regions means they have tight prior uncertainties and hence relatively small anomalies. We performed the same statistical analysis on these series as for the longer ones for larger regions. None of the regions exhibited statistically significant variability under our test, due both to the higher posterior uncertainty (shorter averaging period) and shorter

Figure 4. Five-month triangular-filtered anomalies for the (a) North American, (b) European, and (c) Asian land regions for the CRC-MATCH (black, solid), CCAM (black, dashed), and T3-IAV (red) inversions. The shaded region shows the $\pm\sigma$ uncertainty from the CRC-MATCH inversion. For Asia, the blue dotted line shows the CO_2 (converted to carbon) from the GFEDv2 emissions fields of *Van der Werf et al.* [2006] (updated to the end of 2005).

run length (5 years versus 14). We concentrate here on particular events in the record.

[37] All three regions show considerable variability in this period with, in general, the control inversions being more variable than the T3-IAV mean. North America is the least variable region of the three. This region shows an anomalous

source in 2002 with both models, peaking earlier for CRC-MATCH than CCAM. An anomalous North American source in late 2003 is synchronous for both models while CCAM is slightly earlier for the anomalous sink in late 2004.

[38] The North Asian series is marked by two dramatic excursions, an anomalous sink in mid-2001 (strongest for CCAM) and an anomalous source in mid-2003 (largest for CRC-MATCH). CCAM also shows an anomalous source in late 2004 and early 2005. Note that while the T3-IAV fluxes show a source in mid-2003 the 2001 anomaly is almost absent. Sensitivity experiments neglecting $\delta^{13}\text{C}\text{O}_2$ measurements also produce a weaker sink for CRC-MATCH and CCAM.

[39] Europe shows moderate negative anomalies at the beginning of 2002, mid-2004 and mid-2005 and a large positive anomaly peaking at the beginning of 2003. For all of these CRC-MATCH is more extreme than CCAM.

[40] We have already noted the long-lived northern land anomaly of 2002–2003 (Figure 2). The control inversions suggest that North Asia is responsible for the anomaly in mid-2003. This runs counter to the suggestion of *Ciais et al.* [2005]. This study noted a large drop in productivity at European CO_2 flux measurement sites in the hot and dry summer of 2003. They used modeling studies to upscale this result to a large overall drop in productivity in Europe. This occurred during the period of seasonal drawdown of atmospheric CO_2 so should manifest itself in an anomalous source. Our inversion results show no evidence for this anomaly although the T3-IAV fluxes show a small peak. The inversion does produce a significant anomaly centered in February 2003 for which a biospheric explanation is much more difficult. Note that the 11-month running mean used in Figure 2 takes in contributions from both 2002 and mid-2003.

[41] We have some direct evidence of the location of the mid-2003 anomaly by considering the annual growth rate in CO_2 concentration from various stations (calculated as the difference in concentration of successive Decembers). Six of the 10 stations most closely linked (by atmospheric transport) with the North Asian region occur in the list of 10 highest growth rates for 2003. A potential mechanism for a North Asian anomaly is provided by the Global Fire Emissions Database version 2 (GFEDv2) [*Van der Werf et al.*, 2006]. Figure 4c shows the CO_2 sources from *Van der Werf et al.* [2006] updated to the end of 2005. The sources are spatially integrated and temporally filtered like the inversion fluxes. In 2003 the fluxes of *Van der Werf et al.* [2006] show an anomaly synchronous with the inversion results and with amplitude closer to the CCAM than the CRC-MATCH result. The inversions and GFEDv2 also agree on the timing of an anomaly in 2002. The location of the 2003 fire anomaly is supported by the independent evidence of *Edwards et al.* [2004]. They used observations of the MOPITT instrument to map atmospheric CO concentrations and noted a large positive anomaly in south-eastern Russia (Siberia) in the early summer of 2003. This was also noted in aircraft measurements by *Nedelec et al.* [2005]. The northern land anomaly is longer lived than the

Figure 5. Eleven-month running mean flux anomaly for (a) Europe and (b) southern land for the CRC-MATCH (solid) and CCAM (dashed) inversions with (black) and without (red) $\delta^{13}\text{CO}_2$ data.

North Asian anomaly and we have not found clear explanations for earlier parts of the anomaly.

4. Role of $\delta^{13}\text{CO}_2$ Measurements

[42] In this section we investigate the difficulty and the value of using $\delta^{13}\text{CO}_2$ data. We quantify the difficulty by comparing the quality of fit to CO_2 and $\delta^{13}\text{CO}_2$ data. We quantify the value by considering the impact of $\delta^{13}\text{CO}_2$ measurements on estimated fluxes and their uncertainties.

[43] We first ask how well we can fit the $\delta^{13}\text{CO}_2$ data with the simplified model of atmospheric $\delta^{13}\text{CO}_2$ we use. We apply the algorithm of Michalak *et al.* [2005]. This algorithm calculates a series of multipliers of the data uncertainty such that, in the case of Gaussian residuals, the average quality of fit of the data is consistent with the uncertainties so that $\chi^2 = N/2$ where χ^2 is the cost function from equation (1) and N is the number of observations. The theoretical background is explained by Michalak *et al.* [2005]. Here we naturally subset the data into the CO_2 and $\delta^{13}\text{CO}_2$ data. The required multipliers are 0.87 for the CO_2 and 0.87 for the $\delta^{13}\text{CO}_2$ data. This is consistent with the idea that there are several important processes that affect $\delta^{13}\text{CO}_2$ values in the atmosphere that are not included in our model. We have used the original rather than rescaled

uncertainties throughout this paper. This is a conservative approach since our overall fit is better than our uncertainties would demand. We also note that, using the optimum multipliers, the final cost function is about 20% higher than expected. Presumably the Gaussian assumption is weakly violated for parts of the data.

[44] The parallel inversions performed with and without the inclusion of $\delta^{13}\text{CO}_2$ measurements allow us to assess the impact of these measurements on flux estimates. Figure 5 shows two examples for the southern hemisphere land and for Europe. We show the control inversions and inversions without $\delta^{13}\text{CO}_2$ for both models. For the southern land we see that the choice of model makes less difference than the inclusion of $\delta^{13}\text{CO}_2$ measurements. $\delta^{13}\text{CO}_2$ measurements induce more variability for land and the anticorrelated variability for the ocean noted above. This follows from the relatively strong constraint on the zonal mean source and the ability of $\delta^{13}\text{C}$ to differentiate land and ocean sources. We see below that southern South America is visible to the southern hemisphere $\delta^{13}\text{CO}_2$ measurements which can detect variability in this region. Note that the fluxes estimated without inclusion of $\delta^{13}\text{CO}_2$ are very similar to those from T3-IAV shown in Figure 2e. We see similar behavior for the northern ocean (not shown) with $\delta^{13}\text{CO}_2$ observations increasing variability and increasing the difference with T3-IAV. Increased source resolution also explains some of the difference between all our inversions and T3-IAV. For Europe the position is mixed with the choice of models or the use of $\delta^{13}\text{CO}_2$ both having an effect. $\delta^{13}\text{CO}_2$ data is less able to distinguish terrestrial fluxes at the same latitude than to separate land and ocean fluxes. We do see increased convergence between cases with and without $\delta^{13}\text{CO}_2$ measurements after mid-2004 when the Shetland Islands $\delta^{13}\text{CO}_2$ measurements stopped, so clearly these measurements did have some impact. Part of the difference between Europe and the southern hemisphere is explained by the southern bias of the $\delta^{13}\text{CO}_2$ measurements (see Figure 1b).

[45] Finally we show two measures of the role of $\delta^{13}\text{CO}_2$ in reducing uncertainty on regional fluxes. Figure 6 shows the maximum reduction in posterior uncertainty for each region of the CRC-MATCH inversion with $\delta^{13}\text{CO}_2$ measurements compared to that without. The maximum is taken over all months in 1992–2005. Reductions are generally larger for land regions mainly due to their higher uncertainty. The timing of maximum reductions in adjacent regions depends on seasonal changes in meteorology. Reductions reach 50% for a region in South Asia near the Cape Rama station at which $\delta^{13}\text{CO}_2$ was measured until 2002. Largest reductions usually occur near $\delta^{13}\text{CO}_2$ measurement sites. An exception is southern South America. Sensitivity tests deleting measurement sites suggest this region is constrained by a combination of the southern hemisphere $\delta^{13}\text{CO}_2$ measurements.

[46] Figure 7 shows the uncertainty for the CRC-MATCH land region containing India for inversions with and without $\delta^{13}\text{CO}_2$ measurements. This is the region showing the maximum reduction in Figure 6. We see first a large seasonality in the uncertainty. Cape Rama is subject to a monsoonal circulation for parts of the year so the site varies

Figure 6. Maximum uncertainty reduction (%) in any month from 1992 to 2005 for the CRC-MATCH inversion for the case with $\delta^{13}\text{CO}_2$ data relative to that without. The uncertainty reduction for any given month and region is defined as $100 \times (1 - \sigma^{C^{13}} / \sigma^{\text{CO}_2})$.

strongly between sampling marine and terrestrial air. We see that the two cases generate roughly equal uncertainty during that part of the year when Cape Rama does not act as a constraint. However, once the region is observable by the site the $\delta^{13}\text{CO}_2$ measurements afford a considerable extra constraint. We can compare the $\delta^{13}\text{CO}_2$ data uncertainty to the CO_2 data uncertainty by asking how much CO_2 with the $\delta^{13}\text{CO}_2$ signature of terrestrial CO_2 is required to produce the 0.027‰ average uncertainty in $\delta^{13}\text{CO}_2$ during 1997–1999. This is a measure of the equivalent precision in CO_2 units of the $\delta^{13}\text{CO}_2$ measurements. This is roughly 0.7 ppm, lower than the average CO_2 data uncertainty of 1.3 ppm during the same period. Even if we rescaled the uncertainty as suggested earlier the two uncertainties would remain comparable while $\delta^{13}\text{CO}_2$ measurements retain their capacity to separate land and ocean fluxes.

5. Summary and Conclusions

[47] We have performed a series of inversions using two atmospheric models and a range of data sets, principally the CO_2 data from the GLOBALVIEW- CO_2 product and $\delta^{13}\text{CO}_2$ data from CSIRO-GASLAB. The study is set up mainly to probe the interannual variability of the global carbon cycle for the period 1992–2005. As with previous inversions we see large and statistically significant variability in the tropical land with positive anomalies in 1994, 1997–1998 and 2002. The variations are clearly and significantly correlated with the ENSO index. The supposedly classical relation with tropical ocean anomalies leading but opposing land anomalies is only demonstrated in the largest event in 1997–1998. Ocean variability is centered, as expected [e.g., Feely *et al.*, 1999; Buitenhuis *et al.*, 2006] on the tropical Pacific but the relationship with ENSO is complex.

[48] The analysis is based on statistically significant responses across the whole time series. We also analyzed particular events, realizing the limitations of the approach. The most dramatic events are large excursions in the North Asian flux with anomalous sinks in 2001 and sources in summer 2003. The 2003 North Asian anomaly is sufficient to explain the global growth rate anomaly of 2003 as

opposed to the observed anomaly in primary production over Europe noted by Ciais *et al.* [2005].

[49] This study has taken a different line to most inversions in recent years. Although we have moved toward a higher resolution in source space we have not moved to the pixel-based resolution now common in atmospheric inversions [Rödenbeck *et al.*, 2003b; Peylin *et al.*, 2005b]. Instead we have invested effort in improving the modeling of atmospheric $\delta^{13}\text{CO}_2$. The reduction of uncertainty afforded by these measurements highlights their utility. The work on rescaling of uncertainties suggests caution but the similarity between inversions using or not using $\delta^{13}\text{CO}_2$ measurements reinforces confidence. It is clear that the next major step is an improvement in the modeling of $\delta^{13}\text{CO}_2$. While we have improved the modeling of disequi-

Figure 7. Monthly uncertainties (PgC/yr) for the CRC-MATCH inversions with (dashed) and without (solid) $\delta^{13}\text{CO}_2$ measurements for a land region comprising part of India for the years 1997–1999.

librium fluxes and the spatial variability of fractionation, the temporal variability of both these quantities is still treated simply. Evidence from the uncertainty rescaling calculations suggests we can use the $\delta^{13}\text{CO}_2$ measurements more aggressively once we improve these aspects of modeling. Finally there is a much larger data set of $\delta^{13}\text{CO}_2$ measurements now available and the work of Allison and Francey [2007] has suggested methods for the combination of measurements from different research programmes. We therefore hope to follow this study with one using more detailed modeling of $\delta^{13}\text{CO}_2$ and a wider set of $\delta^{13}\text{CO}_2$ observations. The statistical methods introduced by Michalak *et al.* [2005] and applied here will provide guidance on our ability to use such measurements.

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