

Reconstructing the recent carbon cycle from atmospheric CO₂, $\delta^{13}\text{C}$ and O₂/N₂ observations*

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ABSTRACT

This paper presents an attempt to recover the space–time structure of fluxes of CO₂ to the atmosphere over the period 1980–1995 from atmospheric concentration and isotopic composition measurements. The technique used is Bayesian synthesis inversion in which sources are aggregated into large regions and their strengths adjusted to match observed concentrations. The sources are constrained by prior estimates based on a priori knowledge. The input data are atmospheric CO₂ concentration measurements from the NOAA/CMDL network, ¹³CO₂ composition and O₂/N₂ ratios measured at Cape Grim, Tasmania by CSIRO Atmospheric Research. The primary findings are a relatively large long-term mean ocean uptake of CO₂, and seasonal fluxes over land with similar integrated magnitude, but smaller peak amplitude, compared with those derived by Fung and co-workers. Predicted interannual variability is smaller than reported in previous studies. The largest contributor is the oceanic tropics where fluxes vary on the time scale of the southern oscillation. There is evidence of longer time-scale variation in land uptake. Increases in ocean uptake and northern land uptake in the early 1990s are consistent with a response to the Mt. Pinatubo eruption.

1. Introduction

It is important that we characterize the present distribution of carbon dioxide (CO₂) fluxes to the atmosphere and that the estimates of fluxes be well constrained. Enting et al. (1994) showed that knowledge of the current carbon budget was a strong constraint on future behaviour of the carbon cycle. Since it is the natural response of the global carbon cycle to anthropogenic perturbations which will determine the relationship between emissions and concentrations in future, such constraints have substantial policy importance.

Efforts to characterize the contributions to the

atmospheric carbon budget are beset with difficulties. At the root of these difficulties lies the problem of gross undersampling in both space and time. The spatial undersampling means that any snapshot we make of the budget contains information only at somewhat broad scales (albeit uneven). The more spatially detailed we make our estimates the less precise they become. This problem was demonstrated by Brown (1995) for methane.

In the time domain, the problem is not so much the density of our records as the length of them. The atmospheric carbon budget is subject to substantial interannual variability. One can see such variability in the atmospheric growth rates deduced by Conway et al. (1994) and Keeling et al. (1995). In one sense the interannual variability constitutes noise. In this view we wish to know the long-term average behaviour of the system so

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any shorter term fluctuations should be averaged away. In another sense, the interannual variability should contain valuable information. Understanding the causes of such variability should provide us with valuable information on the processes controlling the carbon budget. It is understanding and, ultimately, predicting these processes which constitutes genuine understanding of the system.

This paper presents an attempt to ascertain both the decadal-scale average of the carbon budget and interannual fluctuations on that average. We do this by combining data on the major fluxes to the atmosphere and the concentration and isotopic composition of atmospheric constituents. Previous studies have generally used subsets of such data. Direct flux measurements, e.g., Wofsy et al. (1993) and Takahashi et al. (1997) are relatively easy to interpret since they measure directly the quantity we wish to know. Flux measurements are expensive, however, so are very sparse. Large-scale averages must be constructed from a few point measurements and detail on interannual variability is rarely available from more than a few sites.

A second approach to characterizing at least the long-term average carbon budget is to study inventory changes which result from fluxes. One may hope that inventory changes, being time integrals of the highly variable fluxes, would produce a more reliable signal. Such calculations have been made by, for example, Kauppi et al. (1992) for European forests, and Quay et al. (1992) for the ocean. The most serious problem with such methods is that the signal, the change in inventory, is very small. The other problem is that inventory changes might also be highly nonuniform, particularly on land and hence difficult to sample.

A third and more common technique is to use atmospheric concentration observations to infer fluxes to and from the atmosphere. This is a classical inverse problem with all the usual difficulties of ill-conditioning. The distribution of surface fluxes contains variability at all spatial and temporal scales while concentration observations are very sparse spatially (perhaps at one hundred sites on the Earth's surface). Of course not all the spatial and temporal detail of the fluxes is interesting for a given scientific question. The relatively diffusive transport in the troposphere mixes or integrates fluxes, thus inevitably losing informa-

tion. Whether the information lost is important to us depends on the question we are investigating.

Also, atmospheric mixing is neither homogeneous nor isotropic. We use a chemical transport model (CTM) to describe this mixing. This has been the most common approach to inferring surface fluxes. Such studies have included Keeling et al. (1989), Tans et al. (1990) and Enting et al. (1995). In general, these studies have sought to infer the fluxes averaged over a few years, with the often implicit assumption that concentrations and fluxes are in equilibrium. At the opposite extreme are the studies of Francey et al. (1995) and Keeling et al. (1995) which seek to infer the time variations of global-scale sources. These studies use isotopic composition of atmospheric CO₂ to infer global totals of fluxes from land and ocean. Our approach combines the two approaches above, seeking to estimate the full space-time structure of sources, at least insofar as computational expense and available observations allow. Such studies have only recently become possible with the availability of long, well-calibrated records with some spatial detail. For another reconstruction of sources from similar data see Law (1999). Previous studies have investigated interannual variability either by studying snapshots meant to represent different times (Ciais et al., 1995; Keeling et al., 1989; Enting et al., 1993) or by studying the time history of zonal mean sources (Conway et al., 1994).

It is important, and difficult, to keep in mind the different quantities inferred by the various techniques. For example, inventory changes tell us what finally comes to reside in a given reservoir. Flux estimates tell us what is passing across a given interface, usually between the atmosphere and surface. Sarmiento and Sundquist (1992) showed that riverine carbon posed an important offset between different kinds of estimates, since it is a flux which will appear in ocean and land inventories but will not be seen in the atmosphere. Unravelling such complexities is beyond the scope of this paper but we will need to touch on them when comparing our proposed budget to others. Similar complexities arise from the comparison of perturbed or total carbon budgets. For example, Sarmiento and Sundquist (1992) suggested that much of the riverine flux formed part of a cycle in the preindustrial carbon budget. The two fluxes to the atmosphere (an outgassing from the ocean

and a sink on land) that closed this cycle complicate interpretation and there is scope for other similar complexity.

In Section 2, we describe the particular features of the methods and data we used in the study. Following this is a presentation of the results then a discussion of these results. We conclude with some potential future directions for such work.

2. Methods and data

Most of the methods and data used in this work have been described in detail elsewhere. In this section we will briefly review those aspects necessary for understanding the later results and concentrate on those features which are novel in this study.

2.1. Synthesis inversion

The technique of synthesis inversion has been widely used in inversion studies using trace gas concentrations. It is described in detail by Enting et al. (1993, 1995). The studies of Keeling et al. (1989) and Tans et al. (1990) for CO₂ and Fung et al. (1991) for methane also use the same technique. Briefly we decompose the surface flux into a set of regions. For each of these regions we prescribe a spatial structure. With each region is associated an unknown source strength. The set of these source strengths comprises the vector of unknowns for which we solve.

We relate the (unknown) source strengths to concentrations through the use of an atmospheric transport model. We calculate the distribution of concentration arising from a source of unit magnitude from each region. We adjust the magnitudes of source strengths until the linear combination of concentrations matches the observations.

In Bayesian synthesis inversion we also specify an initial estimate of the source strengths. The prior estimates serve to stabilize the inversion and to include prior knowledge of sources in the problem. As an example we can use information gained from satellite observations (Fung et al., 1987) as an estimate of the seasonal exchange of CO₂ between the atmosphere and terrestrial biota. Bayesian inversion differs from normal least squares problems in that we minimize a weighted sum of the mismatch between unknowns and

initial guesses as well as predicted and observed data.

This work differs from that of Enting et al. (1995) mainly in the time structure of sources and corresponding changes in the data fitted. Enting et al. (1995) analyzed CO₂ and $\delta^{13}\text{C}$ measurements to produce a steady-state distribution, consisting of a global trend plus an annual mean distribution plus a seasonal cycle for each observing site. The seasonal cycle was expressed in terms of the first two harmonics of a Fourier decomposition of the detrended data. This pattern can be reproduced by an annually repeating set of sources. Thus only one year's sources need be calculated. Also several of the sources have prescribed structure in time, e.g., those representing uptake by the terrestrial biota and biomass burning.

In this work we aim to reproduce the regional fluxes for nearly two decades with a time resolution of a month. Thus we must fit data from our whole target period and allow the sources from each region and each month to vary separately. This represents a great increase in computational expense from a few dozen unknowns to a few thousand. Each source strength corresponds to a square pulse of one month's duration from each region. The response to these pulses consists of a sharply peaked distribution which mixes over time towards a globally uniform distribution. This uniform distribution forms the contribution of the pulse towards the long-term trend.

2.2. Transport model

The atmospheric transport model determines the relationship between the spatial structure of sources and the resulting pattern of concentrations. The veracity of the transport model limits the accuracy with which we can ascertain sources. Extra constraints, such as the need to balance long-term budgets, may limit the impact of transport model errors. Also it is unnecessary to use a transport model accurate on scales smaller than the concentration data can resolve.

The model used in this study is the Goddard Institute of Space Sciences (GISS) Chemical Transport Model. It is the same model used by Enting et al. (1995) and Tans et al. (1990). The GISS model is an off-line CTM with a resolution of $8 \times 10^\circ$ and 9 levels in the vertical. It is driven by 4 hourly subsampled winds from the $4 \times 5^\circ$

GISS GCM. The interhemispheric transport of the CTM has been tuned (Prather et al., 1987) to match the observed gradient of the chlorofluorocarbon CFC-11.

The transport characteristics of the GISS CTM are broadly in line with those of most other models used for CO₂ flux estimates. This convergence was shown by Law et al. (1996) who compared the responses of 12 transport models to sources corresponding to fossil fuel burning and terrestrial biotic exchange. For the fossil fuel case (their Fig. 3) the GISS result lies midway among a cluster of models which contain the majority. This result suggests that the ascription of sources to each hemisphere in an inversion would be similar for most of the models shown.

It is less clear that the GISS model's response to a seasonal source (such as terrestrial biotic exchange) is reliable. The lack of consensus in the response to terrestrial exchange can be seen from Fig. 12 of Law et al. (1996). The annual mean gradient arising from such a purely seasonal source is a result of correlation between seasonal transport and seasonal sources, a covariance investigated by Denning et al. (1995) for one model and demonstrated by Law et al. (1996) for several models and also (although via a different mechanism) by Pearman and Hyson (1986). It is noteworthy that those models which demonstrated large annual mean gradients from the seasonal source contained more comprehensive treatments of the planetary boundary layer (PBL). The relatively unsophisticated treatment of the PBL in the GISS model is one limitation of this study. The number of response patterns we need to calculate limits us to a computationally inexpensive model such as the GISS model. For this series of experiments we need to run 324 different tracers to equilibrium.

The lack of interannual variability in transport in this model is not likely to annul any conclusions we draw regarding interannual variability of sources. The model uses one year of meteorological fields from the parent GCM. There is none of the interannual variability from either internal variability of the atmosphere or responses to external forcing which one would expect from the real atmosphere. Inverse calculations from Law and Simmonds (1996) and forward calculations from Bousquet et al. (1996) suggest that interannual variability of transport plays only a small

role in the interannual variability of observed concentrations or deduced sources.

Our inversion results should be relatively insensitive to the method we use to sample model output for comparison with observations. We use linear interpolation among neighbouring gridpoints to the location of the observing station. The potential problem with this simple approach is that observations are generally biased to represent large-scale (and generally marine) air-masses. Enting et al. (1993) found little sensitivity to the method chosen to calculate responses at sites from model gridpoints. Law (1996) also found little sensitivity to the methods of selection for most aspects of simulated concentration. The importance of selection methods may well increase with increasing resolution since Ramonet and Monfray (1996) noticed considerable sensitivity in their $2.5 \times 2.5^\circ$ model. Our model, even coarser than that used by Law, should not be subject to such sensitivity.

2.3. Sources and prior estimates

As already mentioned, Bayesian synthesis inversion requires us to specify a spatial and temporal pattern for source components. We solve for a scaling magnitude for these source components. We need to specify a prior estimate and uncertainty for these magnitudes. In general, it is preferable to specify too many rather than too few source components and to be conservative regarding prior source uncertainties. These choices mean that the maximum possible information is recovered from the data we fit. The apparently weak constraint given to a larger number of unknowns will be improved if aggregations of sources are considered but it is better to do this after rather than before the inversion. The true constraint on the number of sources fitted is computational since both the cost of calculating response patterns and the cost of resulting matrix decompositions increases with the number of source components.

We must include some source terms which do not directly appear in the atmospheric CO₂ budget. These ancillary sources appear in the budget of $\delta^{13}\text{C}$ in the atmosphere and represent the gross exchange between the atmosphere and either the ocean or the terrestrial biosphere. They occur because of the isotopic disequilibrium between the atmosphere and these surface reservoirs. We also must associate each net source of

CO₂ with its corresponding isotopic signature. Finally we also include the contribution of each source to the atmospheric oxygen budget. The oxygen budget contains only a subset of the terms of the CO₂ budget. Representative equations for each species are shown in Section 7.

We use a temporal resolution of one month for all sources, consistent with our use of monthly averaged data. If we used less temporal resolution (e.g., 6-month triangular pulses) we could not fit the seasonal cycles over land of the prior estimates. The sources have the structure of square pulses, i.e., constant within each month.

While the source components are non-overlapping in time, they are allowed to overlap spatially. This overlap occurs despite the fact that we generally treat ocean and land sources separately. The GISS CTM specifies a land-ocean fraction for

each gridbox so that an oceanic source with uniform flux will vary from gridpoint to gridpoint. Also, a single gridbox may contain a source for both ocean and land.

It is difficult to present all relevant detail of the sources and prior estimates used in a calculation like this. The prior estimates of the annual mean strengths for each source component (except fossil fuel) are shown in Fig. 1a. The prior estimate of a measure of the seasonal amplitude of each source component is shown in Fig. 2a. The isotopic signature and stoichiometric relationship between carbon and oxygen fluxes are shown for each flux in Table 1. Note that for the $\delta^{13}\text{C}$ signatures associated with each source we quote the fractionation relative to the current isotopic signature of the atmosphere. A CO₂ source with the same isotopic signature as the atmosphere would not

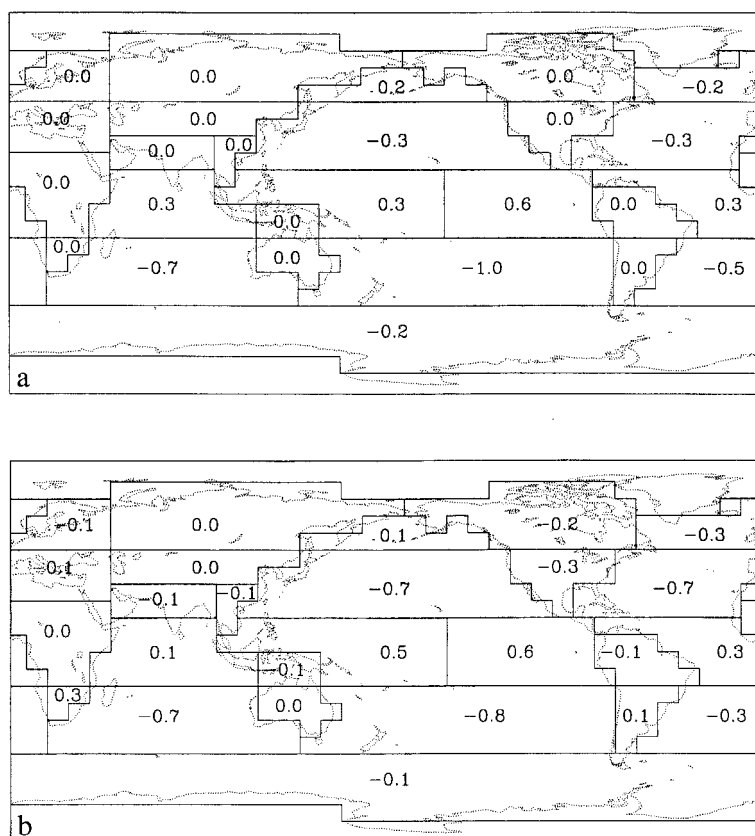


Fig. 1. Prior (a) and predicted (b) annual mean source (Gt C y⁻¹) for our chosen source regions. Region boundaries are denoted by lines. Note that gridboxes may overlap.

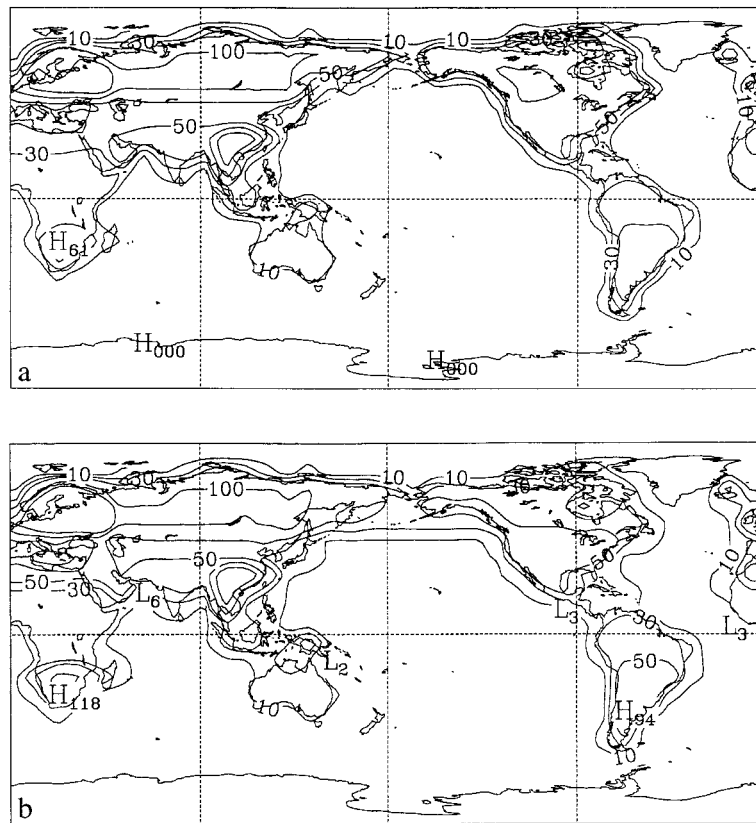


Fig. 2. Prior (a) and predicted (b) growing season net flux. The flux is obtained by averaging all departures of one sign from the annual mean. Contours are at 10, 30, 50, 100 and 150 $\text{gCm}^{-2} \text{yr}^{-1}$.

Table 1. Isotopic signatures and oxygen stoichiometric ratios for the various classes of flux

Flux (Gt C)	Conversion coefficient	
	^{13}C (ppmv ‰)	O_2/N_2 (ppmv)
fossil	−9.51	−0.659
net ocean	−0.683	0
net land	−8.01	−0.495
isoflux	0.471 ^{a)}	0

The values are a conversion coefficient from a unit flux to a unit change in the dependent variables. Note the different units for the data. For ^{13}C , the fractionations are departures from the background isotopic fraction of CO_2 in the atmosphere.

^{a)} The units for the isoflux are $\text{Gt C y}^{-1} \text{‰}$.

contribute to changes in $\delta^{13}\text{C}$. This point is discussed more fully by Enting et al. (1993). Below we describe some of the qualitative features of the sources.

We use one spatial pattern of the source due to fossil fuel release. This is the same as that used by Tans et al. (1990) and Enting et al. (1995) and is based on the work of Marland et al. (1989). The spatial pattern is based on national statistics on fossil fuel burning and cement production and the source is distributed within a country based on population density. There is no seasonal variation in the source. There is an interannual variation reported by Marland and Boden (1997).

The a priori uncertainty for fossil fuel is set at 0.3 Gt C y^{-1} . The carbon isotopic composition of fossil fuel is set at -28.1‰ . The oxygen stoichiometric ratio is set at 1.4 to account for hydrogen

in the fuel. We do not include a source of CO_2 from the oxidation of CO as reported by Enting and Mansbridge (1991). They found including this source made a relatively small difference in their large-scale sources. We should keep the absence of this source in mind when considering uncertainties for large-scale source patterns.

We subdivide the ocean into 12 regions, the same regions as used by Enting et al. (1995). The regions were chosen to match large-scale ocean circulation features and the model grid. Thus each of the major northern hemisphere and southern hemisphere gyres, the tropical upwelling zones, the Antarctic circumpolar regions and the high latitudes of the northern hemisphere are separately represented. We further divide the tropical Pacific into an eastern and western section reflecting the large differences in sea surface temperature between these regions. The regions are shown in Fig. 1.

The prior estimates of ocean flux are based on air-sea $p\text{CO}_2$ measurements summarized in Tans et al. (1990). They show outgassing (positive flux) in the upwelling regions of the tropics and high northern latitudes and uptake in the downwelling regions of the midlatitude gyres. The uncertainties are indicative of the amount of data on the net flux in a particular region. They are generally set at 1.0 Gt C y^{-1} except for the two northernmost boxes which are set at 0.5 Gt C y^{-1} and the South Pacific gyre set at 1.5 Gt C y^{-1} . There is no seasonality in the prior estimates of ocean net flux. The isotopic fractionation of the net flux arises from the mass dependence of solubility. We set this fractionation equal everywhere, following Enting et al. (1995).

We do not associate any oxygen source from the ocean with CO_2 sources from the ocean. This is mainly because oxygen is relatively insoluble in seawater so that changes in atmospheric oxygen content will not equilibrate with the small ocean reservoir. There is a large contribution of ocean biological production to the seasonal cycle of atmospheric oxygen as reported by Keeling and Shertz (1992). Oxygen is used in this calculation as a long-term constraint so seasonal contributions are ignored. We also ignore the possibility of long-term changes in ocean biomass which would involve a net oxygen flux. Some of the subtle difficulties raised by our simple description of the atmospheric oxygen budget will be discussed

when we interpret the long-term partition between land and ocean uptake.

We divide the land into 14 regions as shown in Fig. 1. Unlike the 8 regions used by Enting et al. (1995) our distribution is of contiguous regions rather than representing particular biomes. Our choice of regions is governed partly by the need to represent different biomes but also we need to represent regions with characteristic atmospheric transport influences. As an example, there is an attempt to isolate the regions most strongly affected by the Asian and Australian monsoons. We also consider a net terrestrial flux, rather than separate terms for uptake and release. The separate terms were used by Enting et al. (1995) because of their need to prescribe a time structure for these different types of terrestrial source. We can include this information via the prior estimates of monthly mean flux.

We choose an annual mean terrestrial flux of zero for all regions. This deliberately ignores information from previous studies which seek to deduce the terrestrial flux from budget calculations, e.g., Schimel et al. (1995). To include this information would constitute a double counting of information from global budgets. The large a priori uncertainty in the fluxes allows them to respond freely to the concentration data. We assign an uncertainty of 1.2 Gt C y^{-1} for all net terrestrial sources. The prior estimate of seasonal exchange between the atmosphere and terrestrial biosphere is based on the satellite-derived estimates of Fung et al. (1987). The uncertainty is the same for all months and locations. The $\delta^{13}\text{C}$ of plant material is set at -25‰ for all regions following Enting et al. (1995). This reflects the fractionation that occurs when C_3 plants take up atmospheric CO_2 . Our treatment of $\delta^{13}\text{C}$ means that the global ^{13}C budget does not determine the long-term carbon budget so the choice of fractionation factor is less important. Also, the uniform fractionation does not take into account the varying proportions of C_3 and C_4 plants at different latitudes. We do not use data on spatial gradients of $\delta^{13}\text{C}$ so the error in distributing the source is unlikely to matter.

There is a flux of ^{13}C between the atmosphere and Earth's surface which is not associated with any net CO_2 flux. The flux is usually called the disequilibrium flux or, occasionally, the isoflux. Because we are interested in the CO_2 budget, the isoflux is a so-called nuisance variable. The isoflux

arises from the different isotopic composition of the carbon in the various reservoirs. The isoflux depends on both the difference in composition, i.e., the isotopic disequilibrium and the total amount of CO_2 crossing the interface, i.e., the gross flux. Both terms are highly uncertain. Neither term is of direct interest to us so we carry only the product. The gross flux between the atmosphere and either ocean or terrestrial biota is predominantly controlled by large-scale thermodynamic influences. We expect these influences to change little from year to year. The disequilibrium is increasing because of the input of isotopically depleted CO_2 from fossil fuel. Thus we model the isoflux as the product of an annually repeating gross flux and a linearly increasing disequilibrium. The seasonality of the isoflux is also highly uncertain. The isoflux is represented in our inversion by 12 annually repeating components (i.e., with no variations from year to year) and 12 linearly increasing components where the 12 in each case means one for each month. Note that a linearly increasing source makes a quadratic contribution to the concentration. The choice of parameters means that departures from this quadratic evolution of $\delta^{13}\text{C}$ reflect changes in net fluxes. The isoflux is given large uncertainty which means that $\delta^{13}\text{C}$ data adds little to our knowledge of long-term mean or seasonal fluxes. This arrangement is only possible because of the O_2/N_2 data in this study but other information could also be used to more closely specify the isoflux. Refer to Section 7 for an algebraic description of the ^{13}C source.

2.4. Data

The technique of synthesis inversion is able to use data of a range of types, both different species and different time resolutions. The data for this study range from monthly mean concentrations for CO_2 and values for $\delta^{13}\text{C}$ to a long-term trend in oxygen concentration. Each datum carries both a value and an uncertainty. The uncertainty is the most difficult to assign since it conflates information about measurement error, sampling error and the extent to which the model and the datum represent the same thing.

Our three data sources are CO_2 measurements from the National Oceanic and Atmospheric Administration Climate Monitoring Diagnostics Laboratory (NOAA-CMDL) detailed in Conway

et al. (1994), $\delta^{13}\text{C}$ measurements from the CSIRO GASLAB program detailed in Francey et al. (1995) and the long term oxygen trend from the Cape Grim Air Archive described by Langenfelds et al. (1999). We assume that each of the time series measurements is internally well-calibrated.

The selection of just the Cape Grim $\delta^{13}\text{C}$ data to represent the globe requires comment. Only two records cover the two decades of interest and are described by Keeling et al. (1995) and Francey et al. (1995). There are major differences in interannual behaviour between the two records which are too large to be of biogeochemical origin, and imply systematic error. We note that a third record starting in 1984 by Nakazawa et al. (1997) is in general agreement with both other records after 1988, but registers a significantly larger decrease over the 1986–1988 period than either Keeling et al. or Francey et al. and all three records disagree in 1984. Our preference for the Cape Grim record is based on the very consistent calibration strategy, and support for trends in the 1980s from the Cape Grim Air Archive (Francey et al., 1999).

For all data we assume uncorrelated errors. The error magnitudes are difficult to assign and are adjusted to meet the requirement of statistical consistency with the residuals. The error must account for the presence of unresolved sources (Trappert and Snieder, 1996) as well as the representation of a particular station by a whole gridbox. The assumption of uncorrelated errors has important implications for our predicted uncertainties in fluxes and we will return to this point in the discussion.

The main check on our choice of errors, both for sources and data, is based on the discussion by Tarantola (1987, p. 211). The misfit function should follow a χ^2 distribution with the number of degrees of freedom equal to the number of data. Our large number of data implies that the misfit normalized by initial errors should be very close to 1. We use this requirement to scale both source and data errors. The data uncertainties, after scaling, are 0.85 ppm for CO_2 , 0.012‰ for $\delta^{13}\text{C}$ and 1.1 permeg y^{-1} for the oxygen trend. The CO_2 and $\delta^{13}\text{C}$ uncertainties are based on Enting et al. (1995) and that for oxygen on Langenfelds et al. (1999). For the CO_2 and $\delta^{13}\text{C}$ the uncertainties are much larger than measurement precision. These errors must account for the presence of unresolved sources and the problem of selecting

model and observed concentrations to represent the same thing. The oxygen constraint represents a long-term trend so is not subject to the same problems of representation. We use the error to the fit quoted by Langenfelds et al. (1999) as our base value before scaling.

A major focus of this study is the interannual variability in regional sources. This focus requires a careful choice of the network of observing stations to use. Our general requirement was that data be available for most of the period 1980–1995 and that we would generally not introduce much new data during the inversion. The negative consequence of this is a greatly underused dataset towards the end of the period. To test the importance of new data we have carried out an inversion for the same period using 25 stations. Some of these stations were only available during the 1990s. The network of stations we used is shown in Fig. 3. The sampling is best in the northern hemisphere and poorest over the tropical continents and southern oceans.

3. Results

An inversion with this level of detail produces a substantial body of results. After discussing the quality of fit to the observed data, we will discuss the most useful aggregations of these results. We will move from the most general to the most regionally detailed and from the long-term steady behaviour to interannual variability. Finally we

will discuss the confidence in the predicted sources embodied in the predicted errors.

3.1. Fits to data

The inversion produces a generally good fit to the observed data. This result is not surprising given the excess of source components over data and the weak prior constraints placed on the sources. Two examples of records with different seasonal characteristics are shown in Fig. 4. The fit at South Pole (Fig. 4a) is very good with both seasonal and interannual changes closely matched. The record at Barrow Alaska (Fig. 4b) is less well matched, particularly at the extremes of the seasonal cycle. One reason for this is the time resolution of the sources. The sharp peak in the observed seasonal cycle is produced by sources which vary over less than a month and we cannot resolve this variation.

The ^{13}C and O_2/N_2 observations are also well reproduced by the inversion. The smoothed ^{13}C record from Cape Grim, Tasmania (not shown), is very closely followed. Recall that there are 12 components of gross isotopic exchange which are free to match the mean seasonal cycle in ^{13}C so a good match is hardly surprising. These components can also soak up any error associated with the removal of the seasonal cycle in the data by smoothing. The O_2/N_2 long-term trend is also well matched, -16.4 per meg y^{-1} compared with the observed value of -16.9 ± 0.9 per meg y^{-1} measured by Langenfelds et al. (1999). We will

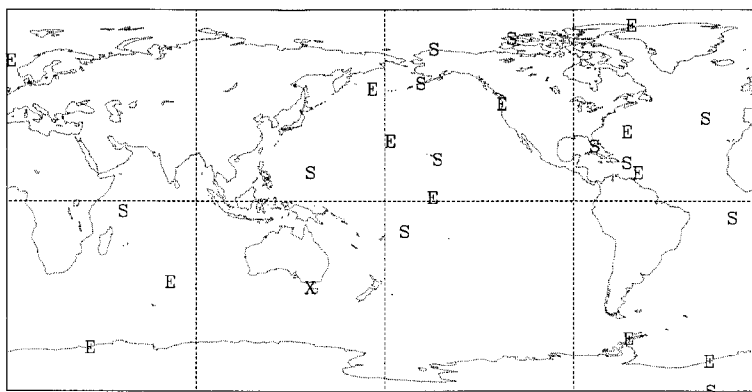


Fig. 3. Locations of monitoring stations supplying observations for the study. The symbol "S" indicates positions of CO_2 monitoring stations for the standard case and "E" for the larger network case. "X" denotes the Cape Grim station which supplied the $\delta^{13}\text{C}$ and O_2/N_2 data for both cases as well as CO_2 data for the extended network case.

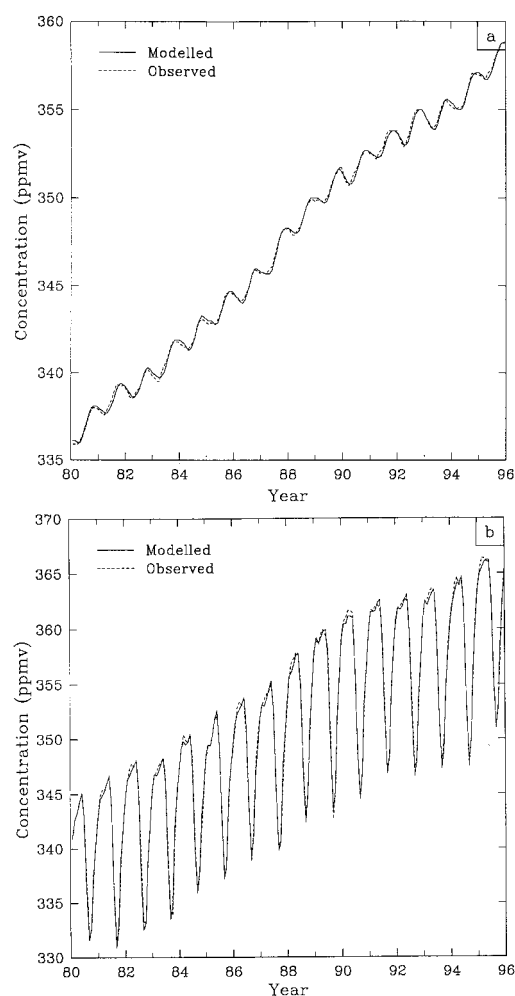


Fig. 4. Modelled and observed CO_2 concentration (ppmv) for South Pole (a) and Barrow Alaska (b) from 1980–1995. The solid line represents modelled concentration while the dashed line represents monthly mean observations taken from the NOAA/CMDL ftp site described by Conway et al. (1994).

discuss the consequences of this slight underestimate when we come to compare our global budgets to others.

3.2. Long-term mean

The long-term mean ocean and land sources are $-2.1 \pm 0.3 \text{ Gt C y}^{-1}$ and $-0.7 \pm 0.3 \text{ Gt C y}^{-1}$ respectively. The overall global partition is deter-

mined by the requirement to match the O_2/N_2 and CO_2 trends.

A representation of predicted annual mean source is shown in Fig. 1b. In general we see relatively small changes in any one region compared to the prior estimates. When aggregated, the predicted fluxes show a shift of uptake from the southern extratropics to the northern extratropics relative to the prior estimate. The southern extratropical uptake falls from 1.8 Gt C y^{-1} to 1.2 Gt C y^{-1} while the northern extratropical uptake increases from 0.4 Gt C y^{-1} to 2.0 Gt C y^{-1} . The tropical source falls from 0.7 Gt C y^{-1} to 0.4 Gt C y^{-1} . There are some differences between land and ocean changes. In general changes in ocean uptake are larger despite the smaller prior uncertainty placed on ocean uptake. The network used for this study is unable to provide strong constraints on the structure of land sources, that is to clearly discriminate between land sources from neighbouring regions. The region with the best constrained land source, North America, does show a larger sink than other land regions.

Fung et al. (1987) defined the growing season net flux (GSNF) as the integrated seasonal flux of one sign between the atmosphere and terrestrial biosphere. For a balanced biosphere we can calculate this flux by averaging all contributions of one sign. For a biosphere with a net annual exchange we first remove the annual mean component so that the GSNF represents only the seasonality. The prior and predicted estimates for this quantity are shown in Fig. 2. We display the GSNF as a flux rather than an integrated total per region to highlight areas of biotic activity. A consequence of this display is that fluxes do not appear to be uniform within regions. This is a consequence of the land-ocean fraction in the underlying transport model.

The prior estimate of GSNF shows two regions of particularly high activity, Western Europe and South-East Asia. The ocean regions are all zero by construction. The general picture for the predicted estimate is of relatively small changes to the prior estimate. The global total changes from 10.1 Gt C y^{-1} to 11.9 Gt C y^{-1} mainly because of the introduction of some seasonality in the ocean. If we considered only the land regions the total changes very little. This might be expected since the same inability of the observing network to

constrain the land annual mean will recur for the seasonal cycle. The hot-spot regions over Eurasia and South-East Asia both intensify slightly, this change being offset by reductions over other northern hemisphere continents.

The long-term, annual mean, zonal mean, non-fossil source is shown as the solid line in Fig. 5. The flux exhibits uptakes in the midlatitudes of both hemispheres and a tropical source. The integrated regional sources are -1.9 Gt C y^{-1} for the northern extratropics (north of 24°N), $+0.4 \text{ Gt C y}^{-1}$ for the tropics (24°S – 24°N) and -1.2 Gt C y^{-1} for the southern extratropics. The dashed line is obtained by a similar inversion on a larger and more variable network, reaching a maximum of 25 stations. The general similarity of the two curves suggests some robustness of the inversion to the choice of network.

The dotted line is a similar quantity from the inversion of Enting et al. (1995), their low growth case. Given the large difference in land-ocean partition of uptake between our study and theirs, the similarity of the zonal means is surprising. There are two main differences. In the southern hemisphere high latitudes, Enting et al. (1995) predict a source while our inversion has a weak sink. On the Equator, Enting et al. suggest a sharp dip in the broad-scale source while ours is much

more uniform. Enting et al. also show more sharply focussed sinks in the extratropics of both hemispheres. Note that the curve of Enting et al. omits their source of CO_2 from CO oxidation in the free atmosphere.

3.3. Interannual variations

One development in this study is the use of time histories of spatially resolved data along with multiple species. To the extent the network can resolve spatial structure, we are in a position to comment on the regional processes which control the observed changes in CO_2 growth rate. For interannual variability, the critical data are the CO_2 time histories from the 12 sites and the $\delta^{13}\text{C}$ record from Cape Grim, Tasmania. The O_2/N_2 trend is included only as a long-term constraint and so does not contribute any information on interannual changes.

The land and ocean fluxes from our calculation are shown as the solid lines in Fig. 6. The dotted lines are the same fluxes from the work of Francey et al. (1995). These global, annual fluxes are directly deducible from the global trends of CO_2 and $\delta^{13}\text{C}$. The long-term mean fluxes are deducible from the CO_2 and O_2/N_2 trends. The CO_2 growth rate is the sum of all source components. Calculating it this way is a less arbitrary process than interpolating concentrations to construct a global mean.

The dominant long-term feature of the time variation in global flux is the large increase in land uptake starting in the late 1980s. This corresponded with a flattening in the $\delta^{13}\text{C}$ trend. In the early 1990s, both uptakes are large, corresponding to a low growth period in CO_2 . The other striking feature on shorter time scales is the increased ocean uptake in 1982–83, 1986–87 and 1991–93. These periods correspond to low growth periods in CO_2 which are not matched by the $\delta^{13}\text{C}$ trends. Similarly, the relatively small ocean uptake in 1984–85 and 1988–89 corresponds to a higher growth rate in CO_2 without a corresponding trend in $\delta^{13}\text{C}$. The land fluxes do not show such strong interannual changes on these time scales, an outcome which cannot be attributed solely to the sparse sampling network.

The interannual changes in global uptake are further disaggregated in Fig. 7. This figure shows the annual mean of the flux for the tropics, north-

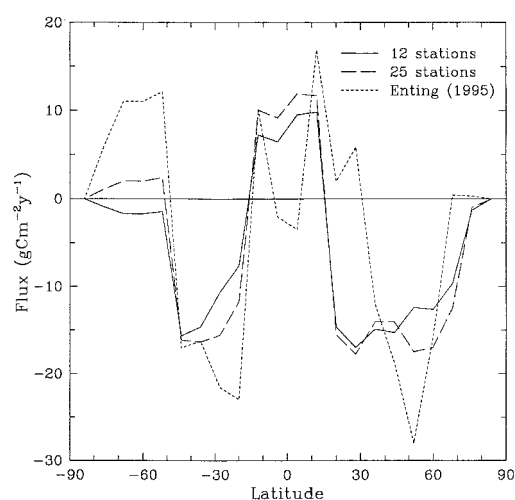


Fig. 5. Long-term, annual mean, zonal mean source ($\text{gCm}^{-2}\text{yr}^{-1}$). The solid line denotes our standard case (12 stations), the dashed line a case with a network increasing to 25 stations and the dotted line the low-growth case from Enting et al. (1995).

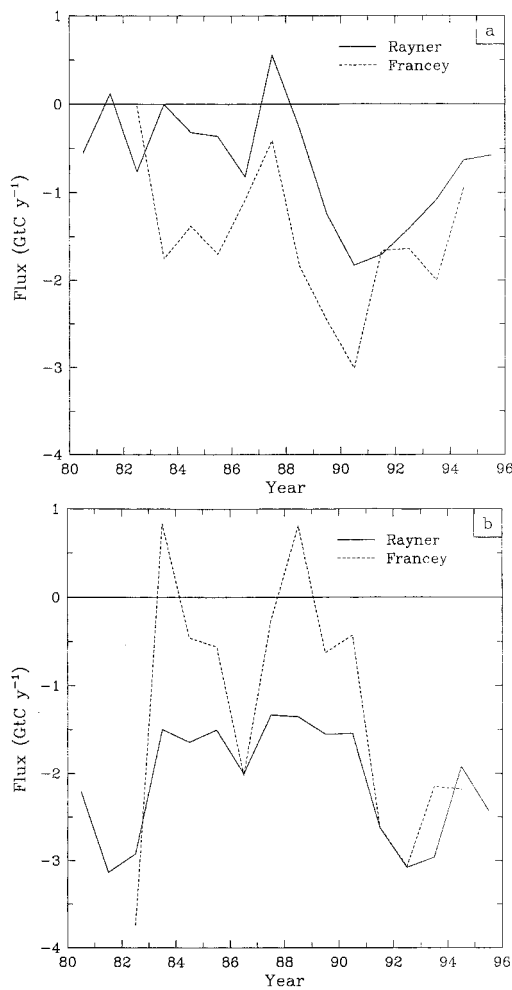


Fig. 6. Annual mean fluxes (GtC y^{-1}) for land (a) and ocean (b). The solid line denotes our inversion while the dotted line denotes the inversion of Francey et al. (1995).

ern extratropics and southern extratropics for land and ocean separately. The fourth curve on each plot is the annual mean southern oscillation index (SOI). The largest contributor to global variability is the tropical ocean. Extratropical variations in oceanic flux, while in phase with the tropical ones, are smaller. Also there is a clear similarity in variations of the tropical source and SOI curves.

The land sources show less short-term variability than the oceans for all regions, a result expected from Fig. 6. The dominant longer-term feature, the increased uptake starting in the late 1980s, commences in the southern extratropics but moves

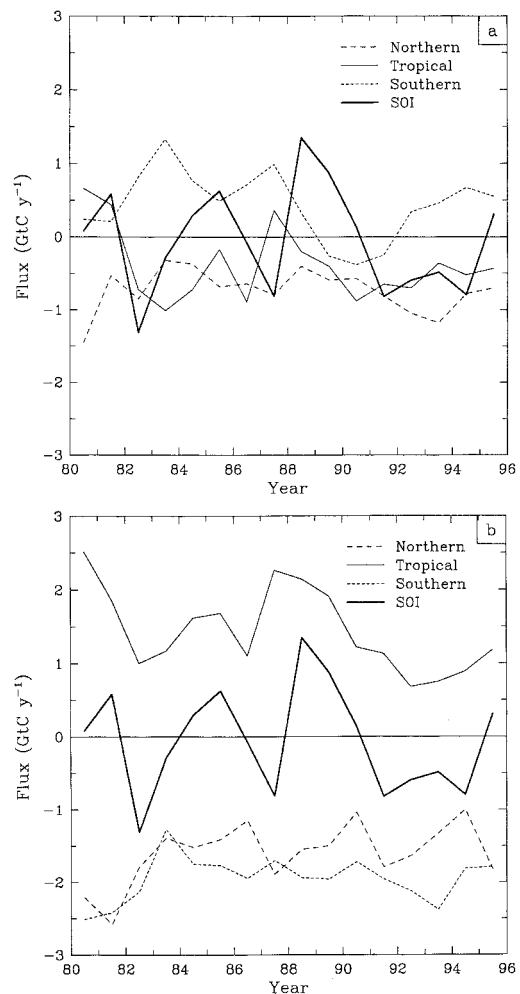


Fig. 7. Annual mean land (a) and ocean (b) fluxes (GtC y^{-1}). The dashed, solid and dotted lines denote northern hemisphere, tropical and southern hemisphere fluxes respectively. The solid, bold line denotes the SOI.

to the northern extratropics in the early 1990s. The relationship with the SOI curve, so evident for the ocean, is apparent for the tropical land although the amplitude of the response is smaller.

4. Discussion

4.1. Long-term mean

The long-term partition between land and ocean fluxes in this study is -2.1 GtC y^{-1} for ocean

and -0.7 Gt C y^{-1} for land. This partition is roughly comparable to other studies using O_2/N_2 and CO_2 data, e.g., Battle et al. (1996) who inferred the trend in O_2/N_2 from air in Antarctic firn. Langenfelds et al. (1999) using a slightly longer record than our study but a single CO_2 site inferred a slightly higher growth rate in CO_2 concentration and a higher ocean uptake and lower land uptake. The CO_2 growth rate is probably more reliable from our inversion, constrained as it is by the records of 12 stations. However our study slightly underestimates the negative oxygen trend, resulting in a larger land uptake. The box model inversion of Langenfelds et al. is required to fit the O_2/N_2 and CO_2 trends exactly. Our ocean uptake is comparable with that from ocean general circulation models, e.g., Sarmiento et al. (1992).

Our ocean uptake is much higher than that from other inversion studies using $\delta^{13}\text{C}$ trends, e.g., Enting et al. (1995), Ciais et al. (1995) and Francey et al. (1995). These differences are large even when one considers the different periods treated by the various studies. The differences appear to lie in the treatment of the gross isotopic flux. For example, the low-growth case of Enting et al. (1995) predicts a global gross isotopic flux of $60 \text{ Gt C y}^{-1} \text{‰}$ for 1986 compared to our value of $81 \text{ Gt C y}^{-1} \text{‰}$ for the same year. A lower gross isotopic flux leads to a higher terrestrial uptake since this is the net flux with the largest isotopic fractionation. Our partition of uptake reflects our view that O_2/N_2 with its simpler budget, represents a more reliable guide to the long-term partition. One caution against this confidence is the finding of Galloway et al. (1995) that nitrogen is being sequestered in the ocean. Any storage of nitrogen in organic form corresponds to a change in carbon storage.

The gross isotopic flux is larger than estimates based on measurements of disequilibria and estimates of gross flux. It should be noted that global aggregation of such sparse measurements is difficult. For the ocean the global flux is a partial cancellation of opposite fluxes since the disequilibrium changes sign from low to high latitudes (Tans et al., 1993). For the land, an estimate of the global isotopic flux requires an estimate of the fluxes from carbon pools of different ages.

Another estimate of the isotopic flux comes from an inversion of the long-term records of CO_2 concentration and $\delta^{13}\text{C}$ from ice-core air bubbles. Such studies are capable of tracking the disequilib-

rium of the ocean and terrestrial biosphere as a function of time. One such study, Trudinger et al. (this issue) predicts an average isotopic flux of $77 \text{ Gt C y}^{-1} \text{‰}$ for the period 1980–1990, very close to our value of $79 \text{ Gt C y}^{-1} \text{‰}$ for the same period. These authors also predict a trend of $1.4 \text{ Gt C y}^{-1} \text{‰ y}^{-1}$ for the same period, comparable to our $1.7 \text{ Gt C y}^{-1} \text{‰ y}^{-1}$.

The spatial distribution of flux within the separate categories of land and ocean surface depends on the characteristics of the transport model used to calculate the response functions and the spatial gradients among the stations of the observing network. This is in contrast with the global partition between land and ocean surface which depends mainly on trends of various species. This study suggests a greater uptake in the northern extratropics than the southern extratropics. This finding is in general agreement with other studies using atmospheric transport models to infer the spatial distribution of fluxes, e.g., Keeling et al. (1989), Tans et al. (1990) and Enting et al. (1995). These studies generally found that interhemispheric atmospheric transport was too weak to reproduce the observed interhemispheric concentration gradient if the dominant sink occurred in the southern hemisphere. The result is more pronounced in the study of Denning et al. (1995) in which the seasonal flux of CO_2 from the terrestrial biota gives rise to an annual mean gradient. This general finding clearly depends on the behaviour of the transport model.

If the atmospheric transport were more vigorous the sink could exist in the southern hemisphere and still produce the observed gradient. In a comparison of atmospheric transport model characteristics, Law et al. (1996) showed that the GISS CTM was intermediate in strength for interhemispheric transport. Thus the predominance of northern over southern sinks seems relatively robust against choice of transport model.

The spatial distribution of fluxes inferred from such an inversion also depends on the network chosen. As we mentioned previously, we chose a relatively sparse network which contained data over most of the study period. To test the sensitivity of the long-term mean to the choice of network we performed another inversion using 25 stations which gradually joined the network during the study period. The extra stations are also shown in Fig. 3. Because these stations were phased in over time, the

average network density lies between 12 and 25 stations. We have already noted the similarity between these curves. There are significant changes at the regional level but it appears that the 12 station network supplies similar information to the 25 station network regarding zonal mean sources. A more detailed study of the resolution properties of various networks is in preparation.

Our study implies a large integrated ocean uptake in the northern extratropics. Given the small ocean areas involved, this corresponds to large fluxes. These fluxes are not, generally, compatible with the air-sea $p\text{CO}_2$ differences reported by Tans et al. (1990). The increase in this region, relative to the prior estimates, is partly offset by a reduction in southern extratropical uptake in the ocean. With the large ocean areas in the southern hemisphere, the flux changes needed to produce these integrated source changes are smaller.

The predicted annual mean land fluxes shown in Fig. 1 are generally small. We have already noted this finding arises from the poor constraint on land sources afforded by the network we used. The 25 station network suggested larger changes.

There is little suggestion, even in the large network experiment, of substantial tropical sources. Recall that we calculate net fluxes. Thus we infer a substantial tropical sink balancing the sources due to land-use change reported by Houghton et al. (1990).

The apparent source over the southern continents is also most likely a function of the network we used. The station density in the southern hemisphere is sufficient to resolve zonal mean sources but rarely the zonal departures needed to distinguish land and ocean sources. Thus, when the inversion requires a decrease in total southern hemisphere sink to match the data both land and ocean display incremental sources. With the ocean starting with a substantial sink this appears as a small change but for the land, with a prior estimate of zero, it produces an apparently surprising source.

We noted in the previous section the relatively small changes in the seasonality of fluxes (expressed as the GSNF) between the predicted and prior estimates. While this apparent lack of response could be blamed on the network used, the same pattern does not hold for another measure of seasonality, the peak to peak amplitude of the flux. This quantity is defined as the difference between the two extreme fluxes in the mean annual

cycle. The peak-to-peak amplitudes (not shown) are substantially altered by the inversion. The general form is of a large reduction in amplitude. The maximum reductions occur for the regions with the largest a priori estimated amplitudes, North America and northern Eurasia. This reduction in amplitude is unlikely to be an artifact of the transport model used in this study. In their comparison of modelled concentrations arising from the same seasonal source fields as we use for our prior estimates, Law et al. (1996) showed that all the models in that study overestimated the peak-to-peak amplitude of seasonal concentration variations at high northern latitudes. This led to the conclusion that the biospheric sources of Fung et al. (1987) may overestimate seasonal amplitude. Our result here supports that conclusion. The suggestion of a broadening in time of the peak source or sink is also consistent with the results of Law et al. (1996) who found that, for example, the draw-down of modelled concentrations started later than the observations but was too vigorous.

4.2. Interannual variability

The interannual variability of global sources for both land and ocean is smaller than either of the box model studies using long records of CO_2 and $\delta^{13}\text{C}$, Francey et al. (1995) and Keeling et al. (1995). In particular our variability in land uptake is roughly half that of Francey et al. and very much less than for Keeling et al. For Francey et al. this is not an artifact of the smoothing used on the $\delta^{13}\text{C}$ record from Cape Grim, since both studies use the same smoothing time of two years. A sensitivity test with a smoothing time of one year did not change the variability much. Nor is it a function of the global CO_2 data used. It is possible that our use of prior estimates reduces interannual variability. Also, our treatment of the Cape Grim $\delta^{13}\text{C}$ record differs from that of Francey et al. They took annual changes at Cape Grim as a reflection of global annual changes. Given the time smoothing on their record this is reasonable. We have monthly changes and a finite transport time from distant locations to Cape Grim. Thus some of the $\delta^{13}\text{C}$ change in a given year will reflect source changes in a previous year. The timing of most of the dominant changes in sources are consistent between this study and that of Francey et al. (1995).

The interannual variability (expressed as the variance of annual means) is similar for our global land and ocean fluxes. This similarity is not, however, a suggestion that their variations are strongly related. The dominant feature in the land fluxes is the large increase in uptake starting in the late 1980s. This has been noted previously by Francey et al. (1995) and is associated with a flattening in the $\delta^{13}\text{C}$ record. The dominant feature of the ocean record is the apparent relationship with the southern oscillation phenomenon previously noted by many authors, e.g., Bacastow (1976), Thompson et al. (1986) and Elliot et al. (1991).

We can shed more light on the features of the global record by looking at the regional contributions in Fig. 7. We have already noted the relative timing of the changes in regional land fluxes. The lead of the southern hemisphere over the other two regions suggests an increase in interhemispheric gradient around 1988. This has been previously noted by Conway et al. (1994). Our inversion interprets this as an increase in the southern hemisphere terrestrial sink earlier than the tropical or northern sink. We are unable to capture the other possible reason for such a change in interhemispheric gradient, a change in transport. Our transport model, with its one year of meteorological data, cannot simulate such a change in transport. The study of Law and Simmonds (1996) suggests that changes in transport do not dominate changes in concentration in such an inversion but nor are they negligible.

The substantial increase in land uptake in the early 1990s is still not sufficient to bring our study into line with those of Keeling et al. (1996) or Ciais et al. (1995). Because these studies consider shorter records, it is difficult to know whether their results represent an anomalous sink (as our study suggests) or a different long-term carbon budget. Under our assumption of poorly known, but constant, isotopic gross fluxes, the potential conflict between the long-term trends in $\delta^{13}\text{C}$ and O_2/N_2 is resolved by tuning these fluxes. It is possible that similar conflicts will arise between inferences about interannual variability derived from O_2/N_2 and $\delta^{13}\text{C}$ records. Such conflicts should contain important information about the variability of the carbon cycle making a joint study using time series of all three species an important future development. The study of

Conway et al. (1994), using only CO_2 but able to recover temporal anomalies, also shows larger anomalous uptake in the early 1990s for the northern extratropics than our study.

The particular similarity noted between tropical sources and the SOI in Fig. 7 suggests that the previously noted global relationships (Bacastow, 1976; Elliot et al., 1991) might well have tropical origins, broadly in agreement with the results of Thompson et al. (1986). Other space–time inversions (Law, 1999; Dargaville, personal communication) have also noted the same relationship. The annual means shown here make it impossible to assess the phase of the relationship or the size of any correlation. We will pursue this topic in more detail in a future study.

The three ocean regions and northern extratropics all show a trend to greater CO_2 uptake starting in 1991. Other authors, e.g., Ciais et al. (1995) and Conway et al. (1994) have invoked the climatic response to the eruption of Mt. Pinatubo in June 1991 to explain this large-scale change in fluxes. Our results are generally consistent with such an explanation. The different responses of tropical and southern land areas remains to be explained.

4.3. Uncertainties

The specification of forms and magnitudes for prior uncertainties is probably the greatest single weakness in this study. The magnitudes were adjusted in order to fit the requirement that the cost function follow a χ^2 distribution. Such a blanket approach is obviously rather rough. The tuning also tells us nothing about the allocation of prior uncertainty among sources and data. We chose to scale the data more than the sources after examining the data residuals but this was a little ad hoc. The large final values for the data uncertainty suggest a large role for unresolved sources and, probably, the need for a more highly resolved set of source components.

The estimated source uncertainty also suggests limitations in our formulation. As might be expected the long-term and global sources are better constrained than regional or annual sources. This follows partly from the need to match long-term or global trends but mainly from the form of the prior uncertainty. The prior uncertainty has the same structure as the predicted uncertainty, the higher the level of aggregation of sources in

either space or time the tighter the confidence interval. This is because the sources are specified with uncorrelated errors so that an average of several will have a smaller error than any of them individually. While such a time structure in the errors is reasonable, the extreme case embodied in our choice (that is, no persistent errors) is unlikely.

The choice for the temporal structure of errors also has implications for the interannual variability of our results. Firstly, correlated errors imply correlated increments so any choice we make about the structure of the prior uncertainty will be directly reflected in the variability of our predicted fluxes. The magnitude of interannual variability we predict is also partly a result of choices about prior uncertainties. In order to allow the long-term mean uncertainty to be large enough it is necessary to make the uncertainties for individual months even larger than they would otherwise be. Future work will address the issue of time-correlated source and data uncertainties.

It is unclear whether the same correlation issues arise spatially. There are two related issues. Firstly we must allow for unresolved source structure within our large regions. We have already discussed our approach to this problem via the method of Trampert and Snieder (1996). It is difficult to quantify the impact of such errors with the tools available to us however the emergence of highly resolved synthesis inversions such as that of Kaminski et al. (1998) will either avoid the issue or at least allow us to assess its importance.

The other problem is the same correlation issue as in the time domain; the relationship between local and larger-scale uncertainties. The problem is less severe in space than in time since the relatively coarse spatial resolution suggests that spatial correlations will be weaker.

4.4. Further work

The comments on uncertainties above suggest some obvious, if difficult improvements which should be made to the work presented here. The technique used in the study is so general and the analysis necessarily so cursory that a series of follow-up studies immediately suggest themselves. Among these are:

- Include a much greater range of data, particularly on spatial gradients of ^{13}C .

- Closer consideration of possible forcing mechanisms for the observed interannual variability. This is already in progress for tropical fluxes.
- Study the use of the technique as a data extension tool.
- Replicate this study for other tracers, particularly methane.

5. Conclusions

This paper has presented an attempt, using synthesis inversion, to infer the time history and spatial structure of CO_2 fluxes to the atmosphere from 1980–1995. The main conclusions can be stated as follows

(1) During this period, the ocean was a much larger sink of atmospheric CO_2 than the land, with an uptake of $2.1 \pm 0.3 \text{ Gt C y}^{-1}$ compared to the land $0.7 \pm 0.3 \text{ Gt C y}^{-1}$.

(2) The major sink occurs in the northern extratropics, with a total sink of 2.0 Gt C y^{-1} north of 25°N . The tropics are a net source of 0.5 Gt C y^{-1} and the southern extratropics a sink of 1.2 Gt C y^{-1} .

(3) The tropical land is a small net sink, suggesting a large biospheric response more than offsetting the source due to land-use change.

(4) Most regional fluxes show large interannual variability on most time scales; the tropical ocean being the most variable.

(5) There is a substantial increase in land uptake through the record, which averages 0.3 Gt C y^{-1} in 1980–1987 and 1.1 Gt C y^{-1} in 1988–1995.

(6) Tropical fluxes show strong variation on the time scale of the southern oscillation suggesting an ENSO-related mechanism for their forcing.

(7) Increases in uptake by the global ocean and northern land after 1991 are consistent with a response to the eruption of Mt. Pinatubo.

6. Acknowledgements

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7. Appendix

Summary equations and computational details

We write the tracer continuity equation schematically as

$$\frac{\partial c}{\partial t} = -\nabla \cdot \Phi + s,$$

where c is concentration, Φ is the flux due to all transport processes and s is source. This equation is linear so we can transform it into an integral Green's function form for the concentration at position $\mathbf{x}$ and time t as

$$c(\mathbf{x}, t) = \int c(\xi, t_0) G_c(\xi, t_0, \mathbf{x}, t) d\xi + \int s(\xi, \tau) G_s(\xi, \tau, \mathbf{x}, t) d\xi d\tau,$$

where t_0 is the initial time, G_c is the Green's function for concentration, and G_s is the Green's function for sources. G_c is sufficiently diffusive that if we consider data only after time $t_1 \gg t_0$ we can replace the first integral by a constant, i.e., a global offset. We need one such offset for each species. We give the offset the initial value of the first datum in the record but a very large uncertainty (Table 2). The delay between the start of the inversion and the first sources that we retain ($t_2 - t_1$) is determined by the atmospheric transport time-scale which, for the globe, is about one year. We allow two years so the initial condition is almost forgotten.

The first data is introduced at t_1 . We allow sources in the period $t_0 - t_1$ although we do not retain them for analysis. These sources serve to match the spatial gradient in the initial data at t_1 so that the gradient need not be established instantaneously by sources at t_1 . We have tested the impact of this spin-up by starting an inversion partway through our study period and confirming that initial sources matched those predicted in the longer inversion.

The main set of decisions in establishing an inversion like this is modelling the source. Below we write the expressions for sources of CO_2 and $\delta^{13}\text{C}$ and the long-term trend in O_2/N_2 . The subscripts f refer to fossil fuel input, l to net land

Table 2. *Uncertainties for prior estimates of sources and data*

Sources	
flux	SD
fossil fuel	0.3 Gt C y^{-1}
north-Atlantic net	0.5 Gt C y^{-1}
north-Pacific net	0.5 Gt C y^{-1}
south-Pacific net	1.5 Gt C y^{-1}
other ocean regions	1.0 Gt C y^{-1}
land regions	1.2 Gt C y^{-1}
fixed isoflux	50 Gt C $\text{y}^{-1} \%$
slope of isoflux	2 Gt C $\text{y}^{-1} \%$ y^{-1}
Data	
data type	SD
CO_2 data	0.7 ppmv
$\delta^{13}\text{C}$ data	3.5 ppmv $\%$
O_2/N_2 trend	0.2 ppmv y^{-1}
initial CO_2	100 ppmv
initial $\delta^{13}\text{C}$	50 ppmv $\%$

These uncertainties reflect the initial choice based on the external sources of information and judgment, before corrections described in the text to ensure consistency between uncertainties and residuals.

sources and o to net ocean sources:

$$S(\text{CO}_2)(\mathbf{x}, t) = a_f(t)S_f(\mathbf{x}) + \sum_1^{12} a_{o,n}(t)S_{o,n}(\mathbf{x}) + \sum_1^{14} a_{l,n}(t)S_{l,n}(\mathbf{x}),$$

where S refers to a prescribed source pattern, and the a_f etc. are the parameters for which we optimize.

For $\delta^{13}\text{C}$, the net sources are weighted by the appropriate fractionation coefficients and a term is added describing the isoflux (see Section 2). Thus

$$S(^{13}\text{C})(\mathbf{x}, t) = f_f a_f(t)S_f(\mathbf{x}) + \sum_1^{12} f_o a_{o,n}(t)S_{o,n}(\mathbf{x}) + \sum_1^{14} f_l a_{l,n}(t)S_{l,n}(\mathbf{x}) + (a_{i,c}(\tau) + a_{i,l}(\tau) \times (t' - t'_0))S_i(\mathbf{x}),$$

where the f terms are the fractionation of net fluxes and $a_{i,c}$ and $a_{i,l}$ are the constant and linearly increasing coefficients for the isoflux. τ represents the month of the year and t' the year. The frac-

tionation factors are quoted in Table 1. Note the units of $S(^{13}\text{C})$ are $\text{Gt C y}^{-1} \text{‰}$.

For O_2/N_2 , the datum represents the long-term trend so that an averaged form of the equation is required. On such long time scales, the spatial structure of the sources is unimportant. Thus we can write

$$T_{\text{ox}} = s_f \overline{a_f(t)} + \sum_1^{14} s_l \overline{a_{l,n}(t)},$$

where T_{ox} is the trend in oxygen concentration (in ppmv) and the s terms are the stoichiometric relations between carbon and oxygen for a particular source. The averaging occurs over time. To summarize the unknowns in the problem: there are source magnitudes from 14 land regions plus 12 ocean regions plus one fossil pattern for each month in 18 years (16 years of data plus two years spin-up), a yearly repeating value and a constant slope for the global isoflux for each month (24 values in all) and offsets for each of the three species in the data. This gives 5859 unknowns in all. The data consist of: (usually incomplete) records of CO_2 concentration from 12 stations from 1980–1995, $\delta^{13}\text{C}$ values for 1980–1995 from

Cape Grim and one value representing the long-term trend in oxygen concentration. In total, this constitutes 2339 data values for our standard case.

The minimization is solved using the usual singular value decomposition (SVD) technique. We divide rows of the Jacobian by the standard deviation for the relevant datum and multiply columns by the standard deviation for the source. A Bayesian singular value λ_b is calculated from the raw singular value as

$$\frac{1}{\lambda_b} = \frac{\lambda}{\lambda^2 + 1}. \quad (1)$$

Since we multiply increments by $1/\lambda_b$ we can see that these increments will be ignored for small λ instead of being amplified as in normal least squares. Thus we do not need to impose a threshold on the singular values.

The SVD is the most computationally demanding part of the inversion. We use the readily available LAPACK routines for the SVD. The standard case takes about two hours on a moderate workstation. The space requirement is set by calculating the a posteriori source covariance. The standard case requires about 53 megawords.

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