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# Terrestrial carbon sink in the Northern Hemisphere estimated from the atmospheric $CO_2$ difference between Mauna Loa and the South Pole since 1959

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ABSTRACT

The difference between Mauna Loa and South Pole atmospheric $CO_2$ concentrations from 1959 to the present scales linearly with $CO_2$ emissions from fossil fuel burning and cement production (together called fossil $CO_2$). An extrapolation to zero fossil $CO_2$ emission has been used to suggest that the atmospheric $CO_2$ concentration at Mauna Loa was 0.8 ppm less than that at the South Pole before the industrial revolution, associated with a northward atmospheric transport of about 1 Gt C $yr^{-1}$ (Keeling et al., 1989a). Mass conservation requires an equal southward transport in the ocean. However, our ocean general circulation and biogeochemistry model predicts a much smaller pre-industrial carbon transport. Here, we present a new analysis of the Mauna Loa and South Pole $CO_2$ data, using a general circulation model and a 2-box model of the atmosphere. It is suggested that the present $CO_2$ difference between Mauna Loa and the South Pole is caused by, in addition to fossil $CO_2$ sources and sinks, a pre-industrial interhemispheric flux of 0.5–0.7 Gt C $yr^{-1}$, and a terrestrial sink of 0.8–1.2 Gt C $yr^{-1}$ in the mid-latitude Northern Hemisphere, balanced by a tropical deforestation source that has been operating continuously in the period from 1959 to the present.

## 1. Introduction

During the last decade, much of the debate over the global carbon cycle has been shaped by two major findings. The first of these is that atmospheric general circulation models (GCMs) constrained by $CO_2$ observations in 1981–1987 suggest that the southward interhemispheric transport of $CO_2$ is comparable to the rate of $CO_2$ accumulation in the atmosphere of the Southern Hemisphere (SH) (Tans et al., 1990). Since the anthropogenic fossil carbon sources are primarily in the Northern Hemisphere (NH), this implies that the net uptake of $CO_2$ by the ocean and land biota in the SH must be small. The anthropogenic $CO_2$ emitted in the mid-latitude NH must therefore be taken up in the same hemisphere (Keeling et al., 1989b; Tans et al., 1990). Using this model transport constraint, and estimates of oceanic uptake from observed air–sea $pCO_2$ gradients, Tans et al. (1990) concluded that the region north of 15°N has an oceanic sink of about 0.6 Gt C $yr^{-1}$ and a land biotic sink of 2–3 Gt C $yr^{-1}$. A large land biotic carbon sink in the mid-latitude NH is supported by measurements of $^{13}CO_2/^{12}CO_2$ and $O_2/N_2$ ratios in the atmosphere (Ciais et al., 1995; Keeling et al., 1996), and by more recent measurements and inverse modeling of atmospheric $CO_2$ (Enting et al., 1995; Fan et al., 1998).

The second major finding is the observation by Keeling et al. (1989a) that the difference in concentrations of atmospheric $CO_2$ between Mauna Loa and the South Pole scales linearly with fossil $CO_2$ emissions from 1959–1988 (data extended to 1994

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in Fig. 1), and that if this linear relation were to be extrapolated to a zero fossil $CO_2$ emission rate, the concentration of $CO_2$ at the South Pole before the industrial era would be higher than that in the NH by 0.8 ppm (Keeling et al., 1989a). This led Keeling et al. (1989b) to hypothesize an atmospheric transport of 1 Gt C $yr^{-1}$ from the SH to the NH before the industrial revolution, balanced by a subsurface oceanic transport from the North Atlantic Deep Water (NADW) formation region to the surface of the Southern Ocean. The pre-industrial carbon transport is of interest because it offsets the oceanic uptake of anthropogenic $CO_2$ in the SH and enhances it in the NH. In order both to satisfy the present atmospheric transport constraint and accomodate the hypothetical pre-industrial carbon transport, Keeling et al. (1989b) proposed that oceanic uptake north of 16°N was 2.3 Gt C $yr^{-1}$ for 1984, which is in significant disagreement with the 0.6 Gt C $yr^{-1}$ uptake for 1981–1987 of Tans et al. (1990).

In support of the hypothetical interhemispheric transport of $CO_2$ by ocean circulation, Broecker and Peng (1992) estimated that the NADW carried about 0.6 Gt C annually to the SH before the industrial revolution. This estimate was based on an analysis of observed concentrations of dissolved inorganic carbon and phosphate. However, the Atlantic transport estimate was later reduced to 0.33 Gt C $yr^{-1}$ in a re-analysis by Keeling and Peng (1995). The Princeton ocean GCM gives a similar Atlantic transport estimate, but the total transport, including the Pacific and Indian Oceans as well as the Atlantic, shows only a small (ca. 0.1 Gt C $yr^{-1}$) southward transport before the industrial revolution (Sarmiento et al., 1995a; Murnane et al., 1999). The ocean GCM also predicts a large (ca. 2 Gt C $yr^{-1}$ in the 1980s) oceanic uptake of anthropogenic $CO_2$ (Murnane et al., 1999).

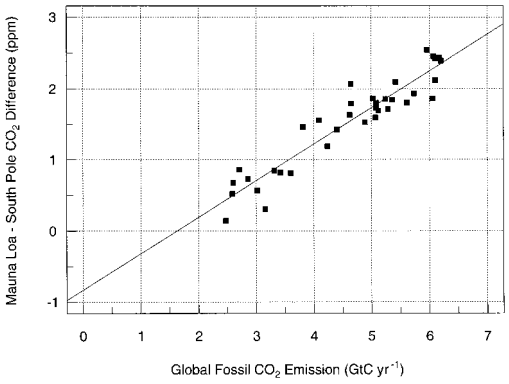


*Fig. 1.* Relationship between the difference of $CO_2$ concentrations between Mauna Loa (MLO) and South Pole (SPO) and the global fossil $CO_2$ emission. Square symbols indicate annual mean data from 1959–1994; the line is a least-squares fit to the data $(C_{MLO} - C_{SPO}) = -0.88 + 0.526E(t)$. Keeling et al. (1989a) examined the linear relationship for the period between 1958–1988.

The purpose of this study is to investigate the inference from Mauna Loa and South Pole $CO_2$ data that there was a large pre-industrial northward $CO_2$ transport in the atmosphere balanced by a southward transport within the ocean. Our analysis employs a 2-box atmospheric transport model to represent the interhemispheric exchange of tracers. We use atmospheric GCMs to determine the relationship between the Mauna Loa to South Pole difference and the difference between the NH and SH mean concentrations.

## 2. Model and data

The simplest model that can address the issue of interhemispheric $CO_2$ exchange is a 2-box model such as that shown in Fig. 2. Fossil $CO_2$ emission is represented explicitly as an input $E(t)$, 95% of which goes into the NH and 5% into the SH (Marland et al., 1994; Andres et al., 1996). Uptake of $CO_2$ by the ocean and land biosphere is represented by $F_N(t)$ for the NH and $F_S(t)$ for the SH. Exchange between the hemispheres is represented as the difference between the hemi-

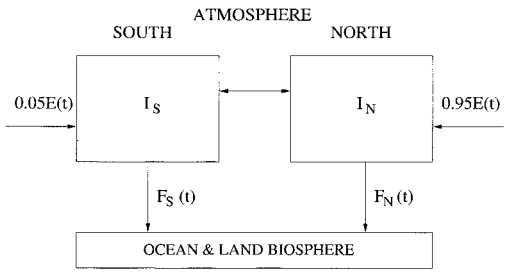


*Fig. 2.* A schematic of the 2-box model. Arrows indicate fluxes of carbon into or out of the atmosphere. $E(t)$ is the rate of global fossil $CO_2$ emission. $F_N(t)$ and $F_S(t)$ are the sum of oceanic and terrestrial uptake of $CO_2$ in the southern and northern hemispheres, respectively.

Table 1. *Model simulated $CO_2$ concentration differences (ppm)*

| $CO_2$ sources | $C_N - C_S$ | $C_{MLO} - C_{SPO}$ | Corrections |
|---|---|---|---|
| fossil emission[a)] | 1.87 | 2.49 | factor, 0.75 |
| terrestrial NEP[b)] | −0.03 | −0.03 | offset, 0.0 |
| air–sea flux[c)] | 0.15 | 0.62 | offset, 0.47 |
| land use flux[d] | −0.04 | −0.11 | offset, −0.07 |

[a)] Model results are shown for the 1990 emissions (Andres et al., 1996). We assume that the ratio of $(C_N - C_S)$ to $(C_{MLO} - C_{SPO})$ varies with time in proportion to fossil $CO_2$ emission.
[b)] This is to correct for the so-called "rectifier effect" due to the coherent variations of transport and the terrestrial net ecosystem productivity (NEP) (Denning et al., 1995).
[c)] The air–sea flux correction is due to the pre-industrial fluxes which we assume continue today (Murnane et al., 1999).
[d)] The land-use term is due to a tropical deforestation rate of 1.6 Gt C $yr^{-1}$ (Houghton and Hackler, 1994).

spheric inventories divided by an exchange time constant: $(I_N - I_S)/\tau$, where $I_N(t)$ is the NH inventory, and $I_S(t)$ the SH inventory. The mass conservation equations for NH and SH inventories are then:

$$\frac{dI_N(t)}{dt} = 0.95E(t) - F_N(t) - \frac{1}{\tau}(I_N - I_s), \tag{1a}$$

$$\frac{dI_S(t)}{dt} = 0.05E(t) - F_S(t) + \frac{1}{\tau}(I_N - I_s). \tag{1b}$$

The problem we are interested in is to determine how the relative magnitudes of $F_N(t)$ and $F_S(t)$ have changed over time, and what this might imply about the pre-industrial interhemispheric exchange of $CO_2$ in the ocean. The assumption is that any pre-industrial interhemispheric exchange of $CO_2$ would have to be due to oceanic processes (though see below for a discussion of the possible role of weathering). The magnitude of such a pre-industrial interhemispheric transport would be $(F_N - F_S)/2$ for any time before the industrial revolution began. We can obtain a solution for the time dependence of this term by subtracting (1b) from (1a) and rearranging the terms:

$$\frac{F_N(t) - F_S(t)}{2} = 0.45E(t) - \frac{1}{2}\frac{d(I_N - I_S)}{dt} - \frac{(I_N - I_S)}{\tau}. \tag{2}$$

We now proceed to determine the magnitude of the terms on the right-hand-side (rhs) of this equation for the period since 1959 for which atmospheric measurements are available.

Historical fossil $CO_2$ emissions are taken from Marland et al. (1994). There are no historical data for $I_N(t)$ and $I_S(t)$. We use the $CO_2$ difference between Mauna Loa and the South Pole $(C_{MLO} - C_{SPO})$, which has been measured since 1959, as a proxy for the difference between NH and SH mean concentrations $(C_N - C_S)$. This is converted to the difference of inventories $(I_N - I_S)$ using a multiplication factor of 1.06 Gt C $ppm^{-1}$. The relationship between $(C_N - C_S)$ and $(C_{MLO} - C_{SPO})$ is estimated by using results from atmospheric GCM simulations for 4 types of carbon sources (Table 1). The bias due to the uneven distribution of fossil $CO_2$ sources is approximately proportional to the global fossil $CO_2$ emission rate, and is corrected by multiplying $(C_{MLO} - C_{SPO})$ by a scaling factor of 0.75 obtained from the GCMs. The "rectifier effect", associated with coherent seasonal variations of atmospheric transport and land biotic metabolism (Denning et al., 1995), causes negligible $CO_2$ differences between the two monitoring stations and between the two hemispheres. The balanced pre-industrial air–sea exchange, with outgassing in the equatorial regions and uptake in the extratropical regions and with minimal interhemispheric transport, causes $(C_N - C_S)$ to be smaller than $(C_{MLO} - C_{SPO})$ by about 0.5 ppm. Land-use changes may have emitted $CO_2$ at a rate of 0.5 to 2 Gt C $yr^{-1}$ in the last few decades, most of which has occurred in the tropics (Houghton and Hackler, 1994). The difference due to tropical deforestation $CO_2$ is remarkably small, about 0.1 ppm in magnitude for a global deforestation source of 1.5 Gt C $yr^{-1}$. In other words, tropical deforestation has only a very small effect on the interhemispheric $CO_2$ difference.

Correcting for the above biases, we have estimated annual mean interhemispheric $CO_2$ differences from 1959–1994 according to $(C_N - C_S) = 0.75\,(C_{MLO} - C_{SPO} - 0.4)$. The ranges in the scaling factor and offset are estimated by comparing model results from 2 different Geophysical Fluid Dynamics Laboratory (GFDL) atmospheric models (GCTM and SKYHI). The scaling factor has a range of about 10%, and the offset has a range of 0.2 ppm. Consideration of other GCMs would undoubtedly increase this range (Law et al., 1996). The resulting $(I_N - I_S)$ is shown in Fig. 3. Data gaps in the monthly average Mauna Loa and South Pole $CO_2$ time series were filled using empirical functions that account for the long-term trends and seasonal variations (Keeling et al., 1989a). Parameters of the empirical functions were estimated based on the monthly $CO_2$ data by the least-squares method.

The atmospheric interhemispheric exchange time has been estimated in a number of studies utilizing surface tracer observations (Jacob et al., 1987; Levin and Hesshaimer, 1996). However, these estimates are applicable only to surface concentrations, not to the 3-dimensional hemispheric mean concentrations in our 2-box model. The range of $\tau$ values given by tracer calibrated GCMs falls between 0.5–1.3 years (Denning et al., 1999). We will use in what follows $\tau = 0.8$ year as calculated in the GFDL GCTM model for tracers such as $SF_6$ and fossil $CO_2$, which are primarily emitted in the mid-latitude NH. The GFDL models predict well the meridional $SF_6$ gradient in the remote marine boundary layer observed by Levin and Hesshaimer (1996).

We are now in a position to estimate $(F_N - F_S)/2$ following eq. (2). The 2nd term on the rhs of eq. (2) is small compared to the 3rd term, and is approximated by the change from times $t-1$ to $t$.

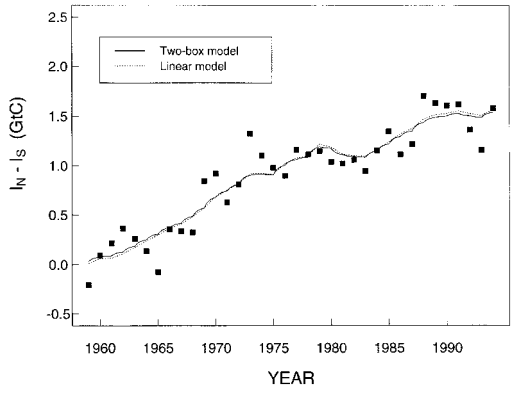


*Fig. 3.* Interhemispheric $CO_2$ difference, $(I_N - I_S)$. The square symbols indicate observational estimates. The solid line shows model results calculated with $\alpha = 0.63$, $F' = 1.2$ Gt C yr$^{-1}$, $\tau = 0.8$ yr, and $E(t)$ from Marland et al. (1994). The dotted line is predicted using the linear model shown in Fig. 1.

## 3. Model result

Fig. 4 shows that $(F_N - F_S)/2$ is of order 0.9 Gt C yr$^{-1}$, with an interannual range of $\sim$0.5 Gt C yr$^{-1}$. The mean value is consistent with Keeling et al. (1989a). There appears to be a slight decrease in the magnitude of $(F_N - F_S)/2$ as $E(t)$ increases, such as would result from an increase with time in the efficiency of SH uptake relative to the NH uptake. If we ignore the interannual variability, the values of $(F_N - F_S)/2$ may be considered linearly, although weakly, related to $E(t)$ for the period from 1959–1994, i.e.,

$$\frac{F_N(t) - F_S(t)}{2} = mE(t) + b, \tag{3}$$

where $m = -0.056 \pm 0.026$ is the slope, and $b = 1.2 \pm 0.1$ Gt C yr$^{-1}$ is the intercept.

Annual growth of atmospheric $CO_2$ has been observed to be linearly related to the global fossil $CO_2$ emissions estimated for the period 1959–1994 (Keeling et al., 1995). On average, 56% of the

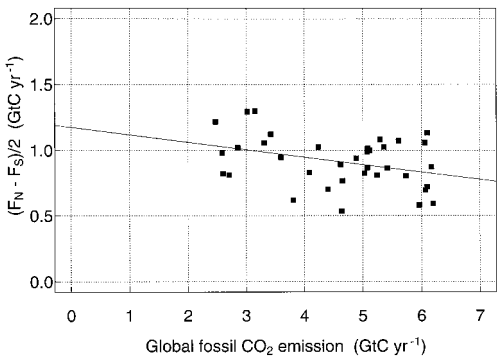


*Fig. 4.* The relation between $(F_N - F_S)/2$ and the global fossil $CO_2$ emissions. The annual change of $(I_N - I_S)$ at $t = 1959$ is assumed equal to that at $t = 1960$ (see eq. (2) in text). The line is a linear least-squares fit to the data.

fossil $CO_2$ remained in the atmosphere. The remaining fraction (44%) must have been taken up by the oceans and land biosphere, i.e.,

$$F_N(t) + F_S(t) = 0.44E(t). \quad (4)$$

Solving eqs. (3) and (4) simultaneously, we obtain,

$$F_S(t) = \alpha(0.44E(t)) - F, \quad (5a)$$

$$F_N(t) = (1-\alpha)(0.44E(t)) + F', \quad (5b)$$

where $\alpha = (0.22 - m)/0.44$ is the fraction of total oceanic and terrestrial uptake of fossil $CO_2$ that occurs in the SH, and $F' = b$ is a constant representing all $CO_2$ sources minus sinks uncorrelated with the fossil emissions from 1959–1994, including land use change emissions, land biotic and oceanic uptake, and the pre-industrial ocean transport postulated by Keeling et al. (1989b). The best fit value for $\alpha$ is $0.63 \pm 0.06$, and for $F'$, $1.2 \pm 0.1$ Gt C yr$^{-1}$. An $\alpha$ of 0.63 implies that 63% of the uptake of fossil $CO_2$ has occurred in the SH. An $F'$ of 1.2 Gt C yr$^{-1}$ represents a SH to NH transport in the atmosphere independent of fossil $CO_2$ emissions. These values are all based on using the GFDL GCTM model (which has a $\tau$ of 0.8 yr) to translate $(C_{MLO} - C_{SPO})$ to $(I_N - I_S)$.

We wish to explore the sensitivity of the "goodness of fit", as represented by the $\chi^2$, to the model parameters. We first combine eqs. (1a), (1b), (5a) and (5b) to obtain

$$\frac{d(I_N - I_S)}{dt} = (0.46 + 0.88\alpha)E(t) - 2F - \frac{2}{\tau}(I_N - I_S). \quad (6)$$

Eq. (6) is integrated in time from 1959–1994 for a range of $\alpha$ and $F'$ values, and for $\tau = 0.8$ yr. Examination of the sensitivity of the parameters to $\tau$ would require comparing results from a range of different atmospheric GCMs with different $\tau$ values, and with their corresponding relationships between $(I_N - I_S)$ and $(C_{MLO} - C_{SPO})$. Our 2 models (GCTM and SKYHI) are too similar to make a useful contrast. Fig. 5 shows the $\chi^2$ as a function of $\alpha$ and $F'$. The ellipses show 68%, 90%, and 99% confidence regions on $\alpha$ and $F'$ jointly, which extend outside the respective confidence intervals of each parameter taken separately. For example, the 68% confidence region corresponds to $\alpha = 0.61 \pm 0.11$ and $F' = 1.1 \pm 0.2$ Gt C yr$^{-1}$.

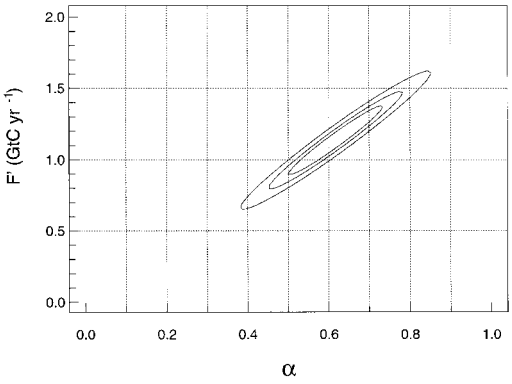


*Fig. 5.* The ellipses are drawn for $\chi^2 = 35.3$, 37.6, and 42.2 (from inside to outside), corresponding to 68%, 90%, and 99% confidence regions, respectively. The $\chi^2$ is calculated with a uniform "measurement noise" that is the standard deviation of residuals for the best fit. The minimum $\chi^2$ is 33.0 located inside the ellipses.

## 4. Discussion

The Princeton ocean biogeochemistry model (OBM) predicts a pre-industrail oceanic transport of 0.12 Gt C yr$^{-1}$ from the NH to the SH (Murnane et al., 1999). A small pre-industrial interhemispheric transport of carbon was also obtained by other ocean GCMs (Stephens et al., 1998; Sarmiento et al., 1999). Here we consider 2 corrections to the ocean model result that increase the pre-industrial transport, and propose a new interpretation to the Mauna Loa and South Pole $CO_2$ observations.

The 1st correction is that the equator is not the appropriate place to draw the boundary of the 2-box model. The atmospheric circulation is divided into southern and northern branches at the inter-tropical convergence zone (ITCZ). The ITCZ varies with season and is not symmetrical about the equator. The OBM is not seasonal, so the best we can do at this time is to consider the annual mean position of the ITCZ. We choose 3°N, the latitude where the annual mean meridional heat transport in the atmosphere goes to zero (Trenberth and Solomon, 1994). Here the interhemispheric ocean transport is 0.29 Gt C yr$^{-1}$.

The 2nd correction is for the southward transport of carbon in the ocean due to weathering and the river input (Sarmiento and Sundquist, 1992; Aumont, 1998). Estimates of the river input

were not available to us at the time the ocean biogeochemistry model was run. Dissolved inorganic and organic carbon is discharged with river water into the oceans, balanced by degassing of $CO_2$ into the atmosphere at steady state. Most of the river input occurs in the NH, but a majority of the degassing occurs in the SH because of the larger ocean area there. This requires a southward ocean transport of carbon. The correction to $F'$ for river input at 3°N is estimated to be 0.2–0.4 Gt C $yr^{-1}$ (Aumont, 1998). Adding the river correction and the correction for the southward transport across 3°N gives a total cross-ITCZ exchange of 0.5–0.7 Gt C $yr^{-1}$.

The remaining difference between $F'$ and the ocean transport is 0.4–0.6 Gt C $yr^{-1}$. We postulate that the remaining difference is associated with terrestrial carbon sources and sinks. This new interpretation involves a coupled tropical deforestation source and a terrestrial sink in the mid-latitude NH. Land use change in tropical Africa and South America is estimated to have emitted $CO_2$ at 0.4 Gt C $yr^{-1}$ in 1960 and 0.9 Gt C $yr^{-1}$ in 1990; tropical Asia emitted 0.3 Gt C $yr^{-1}$ in 1960 and 0.7 Gt C $yr^{-1}$ in 1990 (Houghton and Hackler, 1994). This provides a total tropical source of 0.7 Gt C $yr^{-1}$ in 1960 and 1.6 Gt C $yr^{-1}$ in 1990. But, as noted previously, tropical deforestation sources have almost no effect on the interhemispheric difference of $CO_2$. A way that we can reduce the interhemispheric $CO_2$ difference by the required amount is to have a terrestrial sink in the mid-latitude NH that is twice the required interhemispheric transport of 0.4–0.6 Gt C $yr^{-1}$, i.e., 0.8–1.2 Gt C $yr^{-1}$. In previous model studies, a tropical $CO_2$ source of order $<2$ Gt C $yr^{-1}$ coupled with a land biotic sink of similar magnitude in the mid-latitude NH is found to be consistent with the observed global distribution of $CO_2$ in the early 1980s (Pearman and Hyson, 1986; Tans et al., 1990).

The Mauna Loa and South Pole $CO_2$ data suggest that the NH terrestrial carbon sink may have been in existence, and have been offseting the deforestation source since 1959. This is consistent with analyses based on ocean models and records of $CO_2$ in the atmosphere and in air extracted from ice cores. Previously, a total terrestrial uptake of $1.8 \pm 0.7$ Gt C $yr^{-1}$ is estimated for the period between 1958–1991, which offsets a mean deforestation source of 1.3 Gt C $yr^{-1}$ between 1959 and 1990 (Sarmiento et al., 1995b). Similar results were obtained when ice core and atmospheric measurements of $^{13}CO_2$ and $^{14}CO_2$ were analyzed together with that of total $CO_2$ (Joos and Bruno, 1998). Part of the terrestrial sink must have occurred due to forest regrowth and increased terrestrial productivity in the mid-latitude NH (Dixon et al., 1994; Holland et al., 1997; Randerson et al., 1997; Cao and Woodward, 1998). Carbon sequestration in rice paddies and terrestrial clastic sediments has been estimated to be as much as 1.6 Gt C $yr^{-1}$, most of which occurs in the mid-latitude NH (Stallard, 1998).

There are significant short-term variabilities that are unexplained by fossil $CO_2$ emissions in the 2-box model (Fig. 3), reflecting anomalous oceanic and terrestrial carbon fluxes relative to their long-term trends. The interannual variations in the rate of rise of atmospheric $CO_2$ are discussed by Keeling et al. (1995), who attribute them to anomalous carbon fluxes correlated with global mean surface temperature and precipitation fluctuations. Observations of atmospheric $^{13}CO_2$ have shown evidence for variations in both the oceanic and land biotic uptake of $CO_2$ (Ciais et al., 1995; Francey et al., 1995; Keeling et al., 1995). Interannual variations in atmospheric transport may also cause variations in the interhemispheric difference of atmospheric $CO_2$ concentrations.

## 5. Summary

In summary, we present here an analysis of the difference of atmospheric $CO_2$ concentrations between Mauna Loa and the South Pole from 1959–1994, using a 2-box transport model of the atmosphere. The interhemispheric $CO_2$ difference is related to 2 interhemispheric transport fluxes in the model: (1) fossil $CO_2$ input to the NH atmosphere that is taken up in the SH, and (2) $CO_2$ that is carried within the ocean, or released by the terrestrial biosphere in one region and taken up in another. The latter includes the pre-industrial interhemispheric transport in the ocean that is suggested by Keeling et al. (1989b). The interhemispheric transport flux that is uncorrelated to fossil emissions is estimated to be $1.1 \pm 0.2$ Gt C $yr^{-1}$. However, our ocean GCM predicts a small southward pre-industrial interhemispheric transport of carbon of 0.5–0.7 Gt C $yr^{-1}$ including corrections

for river input and for transport across the ITCZ at 3°N.

We suggest that the remaining interhemispheric transport of 0.4–0.6 Gt C $yr^{-1}$ is due to a modern transport of $CO_2$ from deforestation sources in the tropical region to boreal and temperate terrestrial sinks in the NH, which were present for the entire period between 1959–1994 (not including the effect of interannual variability). Because tropical deforestation causes minimal interhemispheric $CO_2$ difference, the tropical source and the NH sink must each be 0.8–1.2 Gt C $yr^{-1}$ in order to give the same $CO_2$ difference as an interhemispheric exchange of 0.4–0.6 Gt C $yr^{-1}$.

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