

1 **A Joint Global Carbon Inversion System Using Both CO₂ and CO₂ Atmospheric**
2 **Concentration Data**

3

4 Jing M. Chen^{1,2}, Gang Mo², Feng Deng²

5 ¹International Institute of Earth System Science

6 Nanjing University

7 22 Hankou Road, Nanjing, Jiangsu

8 China, 210093

9 and

10 ²Department of Geography and Program in Planning

11 University of Toronto

12 100 St. George Street

13 Toronto, Ontario, Canada M5S 3G3

14

15

16 (Revision submitted to *Geoscientific Model Development*)

17 February 2017

18 Corresponding author: Jing M. Chen, chenj@geog.utoronto.ca

19

20 **Abstract**

21 Observations of $^{13}\text{CO}_2$ at 73 sites compiled in the GLOBALVIEW database are used for an
22 additional constraint in a global atmospheric inversion of the surface CO_2 flux using CO_2
23 observations at 210 sites (62 collocated with $^{13}\text{CO}_2$ sites) for the 2002-2004 period for 39 land
24 regions and 11 ocean regions. This constraint is implemented using prior CO_2 fluxes estimated with a
25 terrestrial ecosystem model and an ocean model. These models simulate $^{13}\text{CO}_2$ discrimination rates
26 of terrestrial photosynthesis and ocean-atmosphere diffusion processes. In both models, the $^{13}\text{CO}_2$
27 disequilibrium between fluxes to and from the atmosphere is considered due to the historical change
28 in atmospheric $^{13}\text{CO}_2$ concentration. This joint inversion system using both $^{13}\text{CO}_2$ and CO_2
29 observations is effectively a double deconvolution system with consideration of the spatial variations
30 of isotopic discrimination and disequilibrium. Compared to the CO_2 -only inversion, this $^{13}\text{CO}_2$
31 constraint on the inversion considerably reduces the total land carbon sink from 3.40 ± 0.84 to
32 $2.53 \pm 0.93 \text{ Pg C y}^{-1}$ but increases the total oceanic carbon sink from 1.48 ± 0.40 to $2.36 \pm 0.49 \text{ Pg C y}^{-1}$.
33 This constraint also changes the spatial distribution of the carbon sink. The largest sink increase
34 occurs in Amazon, while the largest source increases are in southern Africa, and Asia, where CO_2
35 data are sparse. Through a case study, in which the spatial distribution of the annual $^{13}\text{CO}_2$
36 discrimination rate over land is ignored by treating it as a constant at the global average of -14.1% ,
37 the spatial distribution of the inverted CO_2 flux over land was found to be significantly modified (up
38 to 15% for some regions). The uncertainties in our disequilibrium flux estimation are $8.0 \text{ PgC y}^{-1} \%$
39 and $12.7 \text{ Pg C y}^{-1} \%$ for land and ocean, respectively. These uncertainties induced uncertainties of
40 0.47 Pg C y^{-1} and 0.54 Pg C y^{-1} in the inverted CO_2 fluxes for land and ocean, respectively. Our joint
41 inversion system is therefore useful for improving the partitioning between ocean and land sinks and
42 the spatial distribution of the inverted carbon flux.

43

44 **1. Introduction**

45 Over the last few decades, much progress has been made in estimating the global carbon
46 cycle using different methods (*Houghton et al., 2007; Canadell et al., 2007; Le Quéré et al., 2013*).
47 In particular, atmospheric CO₂ mole fractions measured near the surface have been used to infer the
48 carbon flux over land and ocean surfaces through atmospheric inversion (*Rödenbeck et al., 2003;*
49 *Michalak et al., 2005; Peylin et al., 2005; Peters et al., 2007*). However, the uncertainty in the
50 inferred flux is still very large, mostly because of the insufficient number of observation stations and
51 the error in modeling the atmospheric transport of CO₂ from the surface to the observation stations.
52 To reduce this uncertainty, it would be useful to introduce constraints to the inversion using other gas
53 species that are associated the CO₂ flux.

54 Measurements of the atmospheric concentration of the stable isotope ¹³CO₂ at a number of
55 stations across the globe since 1994 have been compiled in a database (*GLOBALVIEW-CO2C13,*
56 *2009*), and the number of extended ¹³CO₂ records from January 1994 to January 2009 increased to 76
57 by 2009. The mole fraction of ¹³CO₂ to CO₂ in the atmosphere is about 1.1%, and the CO₂ exchange
58 between the surface and the atmosphere generally induces concurrent ¹³CO₂ exchange. However, the
59 proportion of the ¹³CO₂ flux relative to the CO₂ flux differs at different locations and different times
60 due to different mechanisms that discriminate against heavier ¹³CO₂ molecules in the exchange
61 processes, and therefore the ¹³CO₂ concentration measured in the atmosphere contains additional
62 information for the CO₂ flux. This information is useful for differentiating between terrestrial and
63 oceanic CO₂ exchanges with the atmosphere because the terrestrial CO₂ flux experiences much
64 greater discrimination against ¹³CO₂ than does the oceanic CO₂ flux (*Tans et al., 1990; Ciais et al.,*

65 1995a; *Francey et al.*, 1995). Observed $^{13}\text{CO}_2$ mole fractions can also provide independent
66 information on the net CO_2 exchange over land and ocean because the net carbon flux to the surface
67 discriminates against heavier $^{13}\text{CO}_2$ (*Fung et al.*, 1997; *Randerson et al.*, 2002; *Suits et al.*, 2005).
68 The $^{13}\text{CO}_2$ observations over the globe, albeit with a limited number of stations, could therefore be
69 used to assist in quantifying the global carbon cycle.

70 In previous studies (*Siegenthaler and Oeschger*, 1987; *Keeling et al.*, 1989a; *Francey et al.*,
71 1995; *Randerson et al.*, 2002), atmospheric $^{13}\text{CO}_2$ observations have been used to separate ocean and
72 land CO_2 fluxes through the use of a technique dubbed “double deconvolution”, by which the CO_2
73 fluxes of land and ocean are separated (deconvolved) based on different discrimination rates against
74 $^{13}\text{CO}_2$ in the atmospheric CO_2 exchange with land and ocean surfaces. This double deconvolution
75 often assumes that the discrimination rates over land and ocean are spatially uniform, although they
76 can be temporally variable. Through forward atmospheric transport modeling, the ocean and land
77 CO_2 fluxes were also separated based on the spatial gradients of the measured $^{13}\text{CO}_2/\text{CO}_2$ ratio either
78 globally (*Keeling et al.*, 1989b) or by latitudinal bands (*Ciais et al.*, 1995a). The same $^{13}\text{CO}_2$ data
79 have also been used in inverse modeling of the surface CO_2 flux (*Enting et al.*, 1995; *Rayner et al.*,
80 1999; *Rayner et al.*, 2008). *Enting et al.* (1995) pioneered a methodology for inverting annual mean
81 ocean and land CO_2 fluxes from both atmospheric CO_2 and $^{13}\text{CO}_2$ concentration data for 12 ocean
82 regions and 8 land ecosystems for the 1986-1987 and 1989-1990 periods. *Rayner et al.* (1999)
83 developed a different methodology to invert monthly CO_2 fluxes for 12 ocean and 14 land regions for
84 the period from 1980 to 1995 from CO_2 observations at 12 stations and $^{13}\text{CO}_2$ and O_2/N_2
85 observations at 1 station. *Rayner et al.* (2008) refined their methodology and applied it to the period
86 from 1992 to 2005 using CO_2 at 67 sites and $^{13}\text{CO}_2$ at 10 sites. These studies showed the usefulness
87 of the additional information from $^{13}\text{CO}_2$ observations in improving the inversion of annual mean and

88 seasonality of the CO₂ flux over land and ocean. In these inversion studies, the discrimination rate for
89 land is either assumed to be a constant (*Enting et al.*, 1995; *Rayner et al.*, 1999) or allowed to vary
90 with the areal fraction of C4 plant in a region (*Rayner et al.*, 2008). These inversions based on the
91 Bayesian principle were also constrained with only simple prior estimates of the terrestrial and
92 oceanic CO₂ and ¹³CO₂ fluxes. Since the data density (the numbers of CO₂ and ¹³CO₂ observation
93 sites) is low, the assumed discrimination constants and these prior estimates would have considerable
94 influence on the inverted results, as this is clearly demonstrated in *Enting et al.* (1995).

95 Atmospheric CO₂ observations have been extensively used to estimate the carbon flux over
96 ocean and land through inverse modeling using Bayesian synthesis (*Gurnay et al.*, 2002; *Rödenbeck*
97 *et al.*, 2003; *Baker et al.*, 2006; *Peylin et al.* 2005) or data assimilation techniques (*Peters et al.*, 2007;
98 *Zhang et al.*, 2014). Atmospheric inversion studies (*Gurnay et al.*, 2003; *Jacobson et al.*, 2007) often
99 produced ocean sinks considerably smaller than those estimated based on observed gradients in
100 dissolved inorganic carbon (DIC) in interior ocean using ocean circulation models (*Steinkamp and*
101 *Gruber*, 2013). Recent estimates for the ocean sink for anthropogenic CO₂ in 2000's based on DIC
102 ranges from 1.6 to 2.6 Pg C y⁻¹ (*Park et al.*, 2010; *Wanninkhof et al.*, 2013; *Landschützer et al.*, 2014;
103 *Majkut et al.*, 2014; *DeVries*, 2014; *Rödenbeck et al.*, 2014) with an uncertainty of about 0.6 Pg C y⁻¹,
104 while atmospheric inversion results are not yet reliable enough to be included in a global ocean sink
105 synthesis (*Le Quéré et al.*, 2013). The partition between ocean and land fluxes using atmospheric
106 inversion techniques is sensitive to errors in atmospheric transport modeling (*Baker et al.*, 2006;
107 *Stephens et al.*, 2007) and prior fluxes for land and ocean used to constrain the inversion (*Zhang et*
108 *al.*, 2014; *Chen et al.*, 2015). It would therefore be highly desirable to use ¹³CO₂ observations to
109 constrain this partition in the inversion process. Accurate partition between ocean and land sinks is
110 important in global carbon cycle research because (1) land sinks are still more reliably estimated as

111 the residual of the global carbon budget than those from land-based data (*Le Quéré et al.*, 2013) and
112 (2) ocean sink estimates based on DIC in ocean water also suffer from considerable errors due to
113 insufficient DIC observations and in ocean circulation modeling (*DeVries*, 2014).

114 The overall goal of this study is to explore the information content of $^{13}\text{CO}_2$ measurements
115 for global CO_2 flux estimation through developing a Bayesian synthesis inversion system that uses
116 both CO_2 and $^{13}\text{CO}_2$ observations. This system is effectively a new double de-convolution system
117 with the capacity to consider the spatial variations of the prior carbon flux and all major isotopic
118 parameters including photosynthetic discrimination, respiratory signature, and disequilibrium rate. In
119 this study, this new system is used to achieve the following objectives: (1) to partition between ocean
120 and land sinks with consideration of the spatial distributions of $^{13}\text{CO}_2$ isotopic parameters over ocean
121 and land; (2) to evaluate the importance of considering the spatial distributions of the $^{13}\text{CO}_2$
122 discrimination rate over land in the inversion of the CO_2 flux, and (3) to assess the impacts of the
123 errors in disequilibrium flux estimation on the flux partition between ocean and land. To achieve
124 these objectives, a terrestrial ecosystem model named the Boreal Ecosystem Productivity Simulator
125 (BEPS) is further developed to simulate the spatial distributions of the $^{13}\text{CO}_2$ discrimination and
126 disequilibrium rates over land for use in a global Bayesian synthesis inversion with $^{13}\text{CO}_2$ constraint.
127 BEPS is also used to produce CO_2 prior fluxes globally to regularize the inversion.

128 **2. Methodology**

129 **2.1 The inversion method**

130 2.1.1 Inversion system

131 The nested inversion system with a focus on North America developed by *Deng et al.* (2007)
132 is adopted in this study. In this system, two of the Transcom regions (*Gurney et al.*, 2002) in North

133 America are divided into 30 regions according to ecosystem types and administrative boundaries
 134 (Figure 1), in order to reduce spatial aggregation errors in the inversion over North America and to
 135 investigate the inverted spatial distribution of the carbon flux against ecosystem model results. This
 136 nested region serves the purpose of evaluating the influence of the spatial distribution of isotopic
 137 discrimination on the inverted carbon flux at a relatively high resolution. Also shown in Figure 1 are
 138 the spatial distributions of 210 CO₂ and 73 ¹³CO₂ observation sites selected in this study from the
 139 NOAA GLOBALVIEW database. Most ¹³CO₂ sites except 11 are collocated with CO₂ sites.

140 2.1.2 Synthesis Bayesian inversion with CO₂ observations

141 To estimate the CO₂ flux (\mathbf{f}), we represent the relationship between CO₂ measurements and
 142 the flux from the surface by a linear model:

$$143 \quad \mathbf{c} = \mathbf{G}\mathbf{f} + \mathbf{A}c_0 + \boldsymbol{\varepsilon} \quad (1)$$

144 where $\mathbf{c}_{m \times 1}$ is a given vector of m CO₂ concentration observations over space and time (m equals
 145 number of stations times number of months, and for CO₂ only inversion, it is 12600, i.e. 210 stations
 146 \times 60 months, 2000-2004); $\boldsymbol{\varepsilon}_{m \times 1}$ is a random error vector with a zero mean and a covariance matrix
 147 $\text{cov}(\boldsymbol{\varepsilon}) = \mathbf{R}_{m \times m}$; $\mathbf{G}_{m \times (n-1)}$ is a matrix representing a transport (observation) operator, where $n-1$ is the
 148 number of fluxes to be determined (equals 3000, i.e. 50 regions \times 60 months, 2000-2004); $\mathbf{A}_{m \times 1}$ is a
 149 unity vector (filled with 1) representing the assumed initial well-mixed atmospheric CO₂
 150 concentrations (c_0) before the first month; and $\mathbf{f}_{(n-1) \times 1}$ is an unknown vector of monthly carbon
 151 fluxes of the 50 regions.

152 Combining matrixes \mathbf{G} and \mathbf{A} as $\mathbf{M}_{m \times n} = (\mathbf{G}, \mathbf{A})$ and vectors \mathbf{f} and c_0 as $\mathbf{s}_{n \times 1} = (\mathbf{f}^T, c_0)^T$, eq.

153 (1) can be expressed as

154 $\mathbf{c} = \mathbf{M}\mathbf{s} + \boldsymbol{\varepsilon}$ (2)

155 The inverse problem of estimating \mathbf{s} from \mathbf{c} is often poorly constrained and a Bayesian
 156 approach is used to circumvent this problem. Pre-existing knowledge and models incorporating
 157 additional sources of information can be used to provide an initial estimate of \mathbf{s} , known as the *a*
 158 *priori*, to constrain the inversion. This *a priori* is then updated when it is combined with information
 159 from \mathbf{c} measurement to form a posterior estimate of \mathbf{s} , known as the *a posteriori*. In Bayesian
 160 synthesis inversion (Tarantola, 1987), the following objective function is employed in the place of
 161 the traditional least square objective function:

162
$$J = \frac{1}{2}(\mathbf{M}\mathbf{s} - \mathbf{c})^T \mathbf{R}^{-1}(\mathbf{M}\mathbf{s} - \mathbf{c}) + \frac{1}{2}(\mathbf{s} - \mathbf{s}_p)^T \mathbf{Q}^{-1}(\mathbf{s} - \mathbf{s}_p)$$
 (3)

163 where $\mathbf{s}_{p \times 1}$ is the *a priori* estimate of \mathbf{s} ; the covariance matrix $\mathbf{Q}_{n \times n}$ represents the uncertainty in the
 164 *a priori* estimate; and $\mathbf{R}_{m \times m}$ is the transport model-data mismatch error covariance. By minimizing
 165 this objective function expressed in eq. (3), we obtain the posterior best estimate of \mathbf{s} as [Enting,
 166 2002]:

167
$$\hat{\mathbf{s}} = (\mathbf{M}^T \mathbf{R}^{-1} \mathbf{M} + \mathbf{Q}^{-1})^{-1} (\mathbf{M}^T \mathbf{R}^{-1} \mathbf{c} + \mathbf{Q}^{-1} \mathbf{s}_p)$$
 (4)

168 Meanwhile the posterior uncertainty matrix for the posterior flux can be deduced as follows:

169
$$\hat{\mathbf{Q}} = (\mathbf{Q}^{-1} + \mathbf{M}^T \mathbf{R}^{-1} \mathbf{M})^{-1}$$
 (5)

170
 171 Following the methodology of Deng and Chen (2011), the CO₂ concentration matrix \mathbf{c} in the
 172 above equations is the residual concentration after subtracting the observed concentration with
 173 contributions from fossil fuel emission, biomass burning, the prior ocean flux and the prior

174 biospheric flux (see Section 2.4 for detail). In this way, the values in \mathbf{s}_p are set to zero and the
175 inverted flux \mathbf{s} is considered to be an adjustment to the prior flux that contributes to the pre-
176 subtracted portions of the CO_2 concentration.

177 **2.1.3 Synthesis Bayesian inversion with both CO_2 and $^{13}\text{CO}_2$ observations**

178 We attempt to use $^{13}\text{CO}_2$ observations to provide an additional constraint to the otherwise
179 CO_2 -only inversion presented above. This additional constraint is possible on the grounds that air
180 $^{13}\text{CO}_2$ concentration is affected differently by carbon fluxes from ocean and land surfaces. Since the
181 $^{13}\text{CO}_2$ gas is transported passively in similar ways as CO_2 , the same transport matrix \mathbf{M} applies to
182 $^{13}\text{CO}_2$ data to associate $^{13}\text{CO}_2$ observations with the surface $^{13}\text{CO}_2$ flux. This simple treatment of the
183 transport matrix differs from Rayner et al. (2008) who considered the reduced response of observed
184 $^{13}\text{CO}_2$ concentrations to surface fluxes with time due to its accumulated exchange with the surface.
185 As we are interested in the net CO_2 flux, the exchanges of both $^{13}\text{CO}_2$ and CO_2 with the surface are
186 consistently not included in the \mathbf{M} matrix calculation, although this simplification would induce
187 errors in the inverted CO_2 flux when the accumulated exchanges are spatially highly heterogeneous.
188 In order to conduct an inversion using both CO_2 and $^{13}\text{CO}_2$ observations, we simply append $^{13}\text{CO}_2$ -
189 related data to the \mathbf{c} , \mathbf{R} and \mathbf{M} matrixes in Eq. (4), while the \mathbf{s} matrix remains unchanged as the
190 purpose of this joint inversion is only to optimize the CO_2 flux. For \mathbf{c} and \mathbf{R} , $^{13}\text{CO}_2$ observations and
191 their variances are appended directly to the original matrixes for the CO_2 only case, as shown in Eq.
192 6. Similarly, the \mathbf{M} matrix is also extended to consider $^{13}\text{CO}_2$ transport, and the relevant elements for
193 the $^{13}\text{CO}_2$ observation stations are from the original \mathbf{M} matrix. However these elements are
194 multiplied by the $^{13}\text{CO}_2$ discrimination rate over land or ocean for each region and each month in
195 order to relate the CO_2 flux to the temporal variations in the measured air $^{13}\text{CO}_2$ composition at each

196 station and each month. The extended \mathbf{M} is a combination of the corrected \mathbf{M} matrix appended to the
 197 \mathbf{M} matrix for CO₂ (see below)

$$\begin{bmatrix} M_{1,1} & M_{1,2} & \dots & \dots & M_{1,n} \\ M_{2,1} & M_{2,2} & \dots & \dots & M_{2,n} \\ \dots & \dots & & & \\ M_{m,1} & M_{m,2} & \dots & \dots & M_{m,n} \\ \hline W_{m+1,1} & W_{m+1,2} & \dots & \dots & W_{m+1,n} \\ \hline W_{m+k,1} & W_{m+k,2} & \dots & \dots & W_{m+k,n} \end{bmatrix} \begin{bmatrix} S_1 \\ S_2 \\ \dots \\ S_n \end{bmatrix} = \begin{bmatrix} C_1 \\ C_2 \\ \dots \\ C_m \\ \hline C_{m+1} \\ \hline C_{m+k} \end{bmatrix}$$

All available CO₂ data

73 stations' ¹³C data

198 (6)

199 where c_i is the CO₂ concentration ($i=1$ to m) and ¹³C composition ($i=m+1$ to $m+k$) in the air from the
 200 starting month ($i=0$); M_{ij} is the transport operator between region-month j (hereafter simply referred
 201 as region) and station-month i (hereafter simply referred as station), and $W_{ij} = D_j M_{ij}$, in which D_j is
 202 the discrimination rate against ¹³CO₂ in the CO₂ flux for region j . In the inversion procedure, the
 203 difference in concentration between two consecutive times is equated with the flux during the time
 204 interval (one month).

205 In order to calculate D_j and C_i ($i=m+1$ to $m+k$) in Eq. 6, some theoretical development is
 206 made according to the ¹³CO₂ budget equation derived by Tans et al. (1993):

$$C_a \frac{d\delta_a}{dt} = F_f(\delta_f - \delta_a) - (F_{lph} - F_{lb})\epsilon_{lph} + F_{lb}(\delta_{lb} - \delta_a) - (F_{oa} - F_{oa})\epsilon_{oa} + F_{oa}(\delta_a^e - \delta_a) \quad (7)$$

208 where C_a is the CO₂ pool in the atmosphere (in Pg C), δ_a is the ¹³C composition of the atmosphere
 209 in ‰, F_f is the carbon emission from fossil fuels and biomass burning, δ_f is the ¹³C composition of
 210 fossil fuels or biomass, F_{lph} is the photosynthetic carbon uptake by the land biosphere (always
 211 positive), F_{lb} is the respiratory carbon flux of the land biosphere (always positive), ϵ_{lph} is the
 212 photosynthetic discrimination of the land biosphere in ‰, δ_{lb} is the ¹³C composition of the land

213 respiratory carbon flux (see Section 2.2.2), δ_{lb}^e is the biospheric ^{13}C composition in equilibrium with
 214 the current atmosphere (i.e. in 2003), F_{oa} is the one-way carbon from the ocean surface to the
 215 atmosphere (always positive), F_{ao} is the one-way carbon flux from the atmosphere to the ocean
 216 surface (always positive), ϵ_{ao} is the air-to-ocean fractionation, ϵ_{oa} is the air-to-ocean fractionation,
 217 and δ_a^e is the ^{13}C composition in equilibrium with the ocean surface. Eq. 7 states that the temporal
 218 variation of the measured ^{13}C composition in the atmospheric CO_2 is determined by contributions
 219 from the various sources: fossil fuels and biomass burning (term 1 of the right hand side of Eq. 7),
 220 net land biosphere carbon uptake (term 2), one-way respiratory flux from the land biosphere (term 3),
 221 net carbon flux of the ocean (term 4), and one-way ocean-to-atmosphere flux (term 5). The one-way
 222 carbon fluxes from land and ocean surfaces are important sources of ^{13}C because the atmosphere is in
 223 isotopic disequilibrium with these surfaces due to the long-term change of the atmospheric ^{13}C
 224 composition. Similar to other terms in Eq. 7, these disequilibrium fluxes are also called isofluxes
 225 (*Rayner, 2001*).

226 In order to reduce the errors of our inversion system (Eq. 6) that assumes linear relationships
 227 between fluxes and concentrations, the contributions of all fluxes, including prior biospheric and
 228 ocean fluxes, to the CO_2 concentration are subtracted from the measured CO_2 concentration prior to
 229 the inversion (*Deng and Chen, 2011*). Accordingly, the contributions of all ^{13}C sources to the ^{13}C
 230 concentration in the atmosphere are also subtracted from the measured ^{13}C concentration. The
 231 purpose of the inversion is then to find the residual CO_2 flux, denoted as \mathbf{S} in Eq. 6. For this purpose,
 232 we denote $S_{IN} = -(F_{lph} - F_{lb})$ as the net flux from the land surface to the atmosphere (negative for sinks)
 233 and $S_{oN} = -(F_{ao} - F_{oa})$ as the net flux from the ocean surface to the atmosphere (negative for sinks).
 234 After taking $S_{IN} = S_{IN}^P + S_l$ and $S_{oN} = S_{oN}^P + S_o$, where S_{IN}^P and S_{oN}^P are the prior net CO_2 fluxes to the

235 land and ocean surfaces, respectively, and S_l and S_o are the residual fluxes to be inverted for the land
 236 and ocean surfaces, respectively, Eq. 7 can be rewritten as:

$$237 \quad S_l \varepsilon_{lph} + S_o \varepsilon_{ao} = C_a \frac{d\delta_a}{dt} - [F_f (\delta_f - \delta_a) + S_{IN}^P \varepsilon_{lph} + F_{lb} (\delta_{lb} - \delta_{lb}^e) + S_{oN}^P \varepsilon_{ao} + F_{oa} (\delta_a^e - \delta_a)] \quad (8)$$

238 Eq. 8 is the theoretical basis for our joint $^{13}\text{C}/^{12}\text{C}$ inversion as it links the measured ^{13}C composition
 239 in the atmosphere to the CO_2 fluxes of the land and ocean surfaces. In the implementation of the joint
 240 inversion system (Eq. 6), a transport matrix is used to link a flux in a particular region to the
 241 concentration measured at a particular site. We focus on optimizing the net CO_2 flux using both CO_2
 242 and $^{13}\text{CO}_2$ observations rather than optimizing the one-way fluxes, and therefore the discrimination
 243 terms to be optimized are moved to the left-hand side of Eq. 8 and the disequilibrium terms remain
 244 on the right-hand side. Based Eq. 8, the regional discrimination D_j in Eq. 6 is therefore defined as:

$$245 \quad \begin{aligned} D_j &= \varepsilon_{lph,j} \quad \text{for land} \\ D_j &= \varepsilon_{ao,j} \quad \text{for ocean} \end{aligned} \quad (9)$$

246 where $\varepsilon_{lph,j}$ and $\varepsilon_{ao,j}$ are the ^{13}C fractionation ratio for region j for land and ocean fluxes,
 247 respectively. In the joint inversion system, we treat S_l and S_o as the state variables and D_j as
 248 predetermined parameters that vary in space (region) and time (monthly). It is therefore prerequisite
 249 to estimate accurately these parameters as well as other isotopic parameters on the right hand side of
 250 Eq. 8.

251 For land regions, BEPS is used to calculate all land variables in Eq. 8, including S_{IN}^P , F_{lb} , ε_{lph} ,
 252 R_{lb} , δ_{lb} and δ_{lb}^e for each region and month. For ocean regions, $\varepsilon_{ao} = -2\text{‰}$, and empirical equations
 253 developed by *Ciais et al.* (1995b) are used to calculate F_{oa} and δ_a^e as functions of sea surface
 254 temperature on $1^\circ \times 1^\circ$ grids.

255 The $^{13}\text{CO}_2$ concentration time series (c_{m+1}, \dots, c_{m+k}) in Eq. 6 in ppm‰ is the numerical
 256 realization of the right hand side of Eq. 8 and is computed with the following equation:

$$257 \quad c_i = \bar{C}_{a,i} \frac{d\delta_{a,i}}{dt} - \sum_{k=1}^5 {}^{13}\delta_k \frac{dC_{k,i}}{dt} \quad (10)$$

258 In Eq.10, $\bar{C}_{a,i} \frac{d\delta_{a,i}}{dt}$ can be calculated with observed CO_2 concentration and ^{13}C composition at two
 259 consecutive times, t and $t+1$, using the following equation:

$$260 \quad \bar{C}_{a,i} \frac{d\delta_{a,i}}{dt} = \frac{C_{a,i}^{t+1} + C_{a,i}^t}{2} (\delta_{a,i}^{t+1} - \delta_{a,i}^t) \quad (11)$$

261 where $\bar{C}_{a,i}$ is the mean concentration of CO_2 at each observation station i between t and $t+1$, and
 262 $\delta_{a,i}$ is the ^{13}C composition at station i , and its derivative with time is taken as its difference between t
 263 and $t+1$. This derivative represents the δ_a growth rate that is the combined outcome of the various
 264 isofluxes in Eq. 7. The term $\sum_{k=1}^5 {}^{13}\delta_k \frac{dC_{k,i}}{dt}$ is the sum of $^{13}\delta$ changes due to fossil fuel and biomass
 265 burning, prior land ^{13}C discrimination flux, land ^{13}C disequilibrium flux, prior ocean ^{13}C
 266 discrimination flux, ocean ^{13}C disequilibrium flux, corresponding to the terms in Eq.8. $^{13}\delta_k$
 267 represents $^{13}\delta$ value (‰) for each term in Eq.8, and $\frac{dC_{k,i}}{dt}$ is the change of concentration (ppm)
 268 calculated with the flux of each term in Eq.8 according to the atmospheric transport function \mathbf{M} in
 269 Eq.6.

270 The uncertainty of c_i as part of the uncertainty matrix \mathbf{R} includes the uncertainties of the six
 271 terms on the right hand side of Eq. 10. The uncertainty for the first term is based on the measurement
 272 error (see next Section 2.1.4) and its global average is 3.08 ppm‰/month. The uncertainties of terms
 273 2 to 6 are estimated to be 0.95, 3.17, 0.87, 0.12, and 2.69 ppm‰/month, respectively. The total

274 uncertainty for c_i is therefore 5.33 ppm%/month as a global average, taking as the square root of the
275 sum of the square of the six uncertainties. As an approximation, this total uncertainty is distributed to
276 each station and each month according to the spatial and temporal patterns of uncertainty of the first
277 term.

278 The inversion system defined by Eq. 6 can be implemented in three ways using (1) CO₂
279 concentration only by excluding the appended matrices for ¹³CO₂, (2) ¹³CO₂ data only by using
280 ¹³CO₂-related matrices only, and (3) both CO₂ and ¹³CO₂ data. Through using the data in these three
281 ways, the information content of ¹³CO₂ measurements for CO₂ can be systematically investigated.

282 In order to investigate the influences of the isotopic discrimination and disequilibrium over
283 land and ocean on the inversion results, we conduct five sets of inversions for the following cases:
284 Case I: The spatial variations of all isotopic compositions and the discrimination and disequilibrium
285 fluxes in Eq. 8 are considered for both land and ocean. This is the ideal case as the basis to
286 investigate other cases; Case II: The photosynthetic discrimination (ε_{ph}) over land is taken as a
287 constant of -14.1‰, which is the global average obtained by BEPS, and therefore $D_j = -14.1\%$. This is
288 a case to ignore regional differences in isotopic discrimination over land; Case III: All isotopic
289 variables are the same as Case I, but the land disequilibrium term in Eq. 8 is ignored. This is a case to
290 investigate the influence of the land isotopic disequilibrium on the CO₂ flux inversion; Case IV: All
291 isotopic variables are the same as Case I, but the ocean disequilibrium term in Eq. 8 is ignored. This
292 is a case to investigate the influence of the ocean isotopic disequilibrium on the CO₂ flux inversion;
293 and Case V: Both land and ocean disequilibrium terms are ignored, but all other isotopic variables in
294 Eq. 8 are same as Case I. This is a case to investigate the importance of the total disequilibrium flux
295 in CO₂ flux inversion at the global scale. Cases III to V are useful not only for evaluating the

296 performance of the joint inversion system but also for assessing the impacts of errors in isotopic
297 disequilibrium estimation on the CO₂ flux inversion.

298 **2.1.4 Covariance matrixes for the CO₂ flux and CO₂ and ¹³CO₂ concentration** 299 **measurements**

300 In the joint inversion using both CO₂ and ¹³CO₂ measurements, the covariance matrix (**Q**) for
301 the CO₂ flux remains the same as that in the CO₂ only inversion (Eq. 3) but the error matrix (**R**) for
302 concentration measurements is expanded to the dimension of 16980×16980 to include 60 months of
303 ¹³CO₂ observations at 73 stations. Following *Deng and Chen (2011)*, we use an uncertainty of 2.0 Pg
304 C y⁻¹ for the total global land surface CO₂ flux, and this total uncertainty is spatially distributed to the
305 39 regions according to the annual total NPP of these regions simulated by BEPS. For each region,
306 the annual total uncertainty is further distributed to each month according to the simulated seasonal
307 variation in NPP. The global total uncertainty (standard deviation) is spatially and temporally
308 distributed in such a way that the total variance is preserved after the distributions, following the
309 principle of TRANSCOM 3 (Gurnay et al., 2003). The uncertainty for the total ocean flux is
310 prescribed as 0.67 Pg C y⁻¹ (*Deng and Chen, 2011*). In this way, all the diagonal elements (Q_{ii}) in the
311 uncertainty matrix **Q** are determined, while off-diagonal values are assigned to zero, meaning that no
312 flux covariances between regions and months are assumed. The uncertainty of CO₂ measurements in
313 the **R** matrix is the same as that described in *Deng and Chen (2011)*, following the approach of
314 *Peters et al. (2005)* and *Bakers et al. (2006)*. In this approach, the uncertainty of a monthly CO₂
315 measurement at a site is estimated as $R_{ii} = \sigma_{const}^2 + GVsd^2$, where constant portion σ_{const} in ppm is
316 assigned according to site category: Antarctic (0.15), oceanic (0.30), land and tower (1.25),
317 mountain (0.90), and aircraft (0.75), while the site-specific variable portion $GVsd$ is obtained from

318 the GLOBALVIEW-CO2 2008 database. The $^{13}\text{CO}_2$ measurement uncertainty is calculated in a
319 similar way: the variable portion is obtained from the GLOBALVIEW-13CO2 2008 database, while
320 the constant portion is taken as $R_a \sigma_{const}$ in ppm first, where R_a is the ratio of $^{13}\text{CO}_2$ to CO_2 in the air
321 (~ 0.011147), and then converted to ‰. The average standard deviation of $\delta^{13}\text{C}$ observations
322 determined in this way for 73 stations is 0.0685‰.

323

324 **2.2 Prior CO_2 and $^{13}\text{CO}_2$ flux estimation**

325 **2.2.1 CO_2 flux**

326 **Terrestrial biosphere fluxes**

327 A process-based terrestrial ecosystem model called the Boreal Ecosystem Productivity Simulator
328 (BEPS) (*Chen et al.*, 1999; *Liu et al.*, 1997) is used in this study to estimate the net terrestrial CO_2
329 flux and its components including the gross primary productivity (GPP), net primary productivity
330 (NPP), heterotrophic respiration (F_{lb}), and net ecosystem productivity (NEP). GPP is calculated using
331 the Farquhar's leaf-level model (*Farquhar et al.*, 1980) upscaled to the canopy level using a recently
332 refined two-leaf approach (*Chen et al.*, 2012). NPP is taken as 45% of GPP (*Ise et al.*, 2010) as
333 global biomass data and its components (stem, foliage, root) are lacking for reliable computation of
334 the autotrophic respiration. F_{lb} is calculated as the sum of the decompositional CO_2 release from 9
335 soil carbon pools, namely coarse and dead wood detritus pool, surface structural pool, surface
336 metabolic pool, surface microbial pool, fine-root structural litter pool, fine-root metabolic pool, soil
337 microbial pool, slow carbon pool, and passive carbon pool. The sizes of these pools for each cover
338 type in each 1° grid are estimated using a model spin-up approach based on simulated NPP in 2000 to
339 create a global land sink of 3.73 Pg C y^{-1} . The total NPP for each 1° grid is taken as a weighted sum

340 of NPP of 7 aggregated land cover types, and the weights are proportional to the areal fractions of the
341 cover types determined using the GLC2000 land cover map at 1 km resolution (*Chen et al.*, 2012).
342 Remotely sensed LAI [*Deng et al.*, 2006] at 1 km resolution and a clumping index map at 6 km
343 resolution (*Chen et al.*, 2005) and a soil textural map (*Webb et al.*, 1991) are aggregated to 1° grids
344 for each cover type based on GLC2000 land cover and used as input to BEPS. National Center of
345 Environmental Prediction (NCEP) reanalyzed data [*Kalnay et al.*, 1996; *Kanamitsu et al.*, 2002] are
346 the meteorological drivers for BEPS to simulate hourly carbon fluxes. The output of BEPS used as
347 the prior flux in the inversions is NEP, which does not include carbon emission due to disturbance.

348 **Ocean fluxes**

349 The daily flux of CO₂ across the air-water interface used in this study is constructed based on
350 the results of daily CO₂ fluxes simulated by the OPA-PISCES-T model [*Buitenhuis et al.*, 2006].
351 This model is a global ocean general circulation model (OPA) [*Madec et al.*, 1998] coupled to an
352 ocean biogeochemistry model (PISCES-T) [*Aumont et al.*, 2003; *Buitenhuis et al.*, 2006]. PISCES-T
353 represents the full cycles of C, O₂, P, Si, total alkalinity and a simplified Fe cycle. It also includes a
354 representation of two phytoplankton, two zooplankton and three types of dead organic particles of
355 different sinking rates. OPA-PISCES-T is forced by daily wind stress and heat and water fluxes from
356 the NCEP reanalyzed data [*Kalnay et al.*, 1996, *Kanamitsu et al.*, 2002]. Hourly S_o (¹³C) is calculated
357 with gridded optimum interpolation sea surface temperature of NOAA National Climate Data Center
358 (*Reynolds and Smith*, 1994; *Reynolds et al.*, 2002).

359 **Fossil-fuel emissions**

360 The fossil fuel emission field (2000-2004) used in this study (<http://carbontracker.noaa.gov>)
361 is constructed based on (1) the global, regional and national fossil-fuel CO₂ emission inventory from
362 1871 to 2006 (CDIAC) [*Marland et al.*, 2009], and (2) the EDGAR 4 database for the global annual

363 CO₂ emission on a 1° grid [*Olivier et al.*, 2005]. The ¹³CO₂ flux from fossil-fuel consumption is
 364 calculated from CO₂ emissions of different fuel types multiplied by their respective ¹³C/¹²C ratios
 365 with consideration of their latitudinal distributions based on *Andres et al.* (2000).

366 **Fire emissions**

367 CO₂ emissions due to vegetation fires are an important part of the carbon cycle [*van der Werf*
 368 *et al.*, 2006]. Each year, vegetation fires emitted around or more than 2 PgC of CO₂ into the
 369 atmosphere, mostly in the tropics. The fire emission field used in this study is based on the Global
 370 Emissions Fire Database version 2 (GFEDv2) (*Randerson et al.*, 2007; *van der Werf et al.*, 2006)

371 **2.2.2 ¹³CO₂ flux**

372 Based on the initial work of *Chen et al.* (2006), BEPS is further developed to include a
 373 capacity to compute the global distribution of the terrestrial ¹³CO₂ flux. Following the principle of
 374 multi-stage ¹³C fractionation in the pathway through leaf boundary layer, stomates, mesophyll and
 375 chloroplast initially proposed by *Farquhar et al.* (1984, 1989) and implemented globally by *Suits et*
 376 *al.* (2005), we developed a module in BEPS for computing the total photosynthetic fractionation and
 377 the resultant ¹³CO₂ flux. Specifically, the photosynthetic discrimination for C3 plants (Δ_{PC3}) is
 378 calculated from

$$379 \quad \Delta_{PC3} = \frac{pA}{C_a} \left[\frac{\Delta_b}{g_b} + \frac{\Delta_s}{g_s} + \frac{\Delta_{diss} + \Delta_{aq}}{g_m} \right] + \frac{C_c}{C_a} \Delta_f \quad (12)$$

380 where Δ_b , Δ_s , Δ_{diss} , Δ_{aq} , and Δ_f are the rates of discrimination against ¹³CO₂ through leaf boundary
 381 layer, stomates, dissolution in mesophyll water, transport in aqueous phase, and fixation in
 382 chloroplast, respectively, and are assigned values of 2.9‰, 4.4‰, 1.1‰, 0.7‰ and 28.2‰,
 383 respectively (*Suits et al.*, 2005). A is the photosynthetic rate in mol m⁻² s⁻¹ and p equals to
 384 $0.022624T_a/(273.16P)$ with the dimension of m³mol⁻¹, where T_a is air temperature in °K and P is the

385 standard air pressure at 1.013Bar. C_a and C_c are the CO₂ concentrations in mol mol⁻¹ in the free air
 386 and leaf chloroplast, respectively. For C4 plants, the photosynthetic discrimination (Δ_{PC4}) is taken as
 387 a constant of 4.4‰ (Suits *et al.*, 2005).

388 The leaf boundary-layer (g_b) is calculated with the following equation

$$389 \quad g_b = \frac{\alpha N}{0.5l} \quad (13)$$

390 where α is the diffusivity of CO₂ in dry air in m²s⁻¹ calculated as $10^{-6}(0.129+0.007T_a)$ and T_a is the
 391 air temperature in °C; l is the leaf characteristic dimension in m, taken as a constant of 0.1 m; and N
 392 is the Nusselt number equal to $(u_d l / \nu)^{0.5}$, where u_d is the wind speed in m s⁻¹ at the vegetation
 393 displacement height (80% of the average vegetation height) and ν is the kinematic viscosity of dry air
 394 in m² s⁻¹ calculated as $10^{-6}(0.133+0.007T_a)$. u_d is derived from the wind speed above the canopy
 395 based on LAI and vegetation height assigned according to plant functional type (Table 1).

396 As part of the GPP calculation, the stomatal conductance (g_s) computed separately for sunlit
 397 and shaded leaves using the Ball-Berry equation (Ball, 1988),

$$398 \quad g_s = f_w \left(m \frac{A h_s}{C_s} p + b \right) \quad (14)$$

399 where f_w is a scaling factor depending on soil moisture and texture (Chen *et al.*, 2012); h_s is the air
 400 humidity at the leaf surface; C_s is the CO₂ concentration at the leaf surface; p is the same as in Eq. 12;
 401 and m and b are the slope and intercept in this linear relationship, and they are assigned values
 402 according to plant function type (Table 1) (Chen *et al.*, 2012).

403 The mesophyll conductance g_m is calculated based on the method of Harley (1992):

$$404 \quad g_m = \frac{A}{C_i - \frac{\Gamma \cdot [J + 8 \cdot (A + R_d)]}{J - 4 \cdot (A + R_d)}} \quad (15)$$

405 where A is the photosynthetic CO_2 assimilation rate; C_i is partial pressure of CO_2 in the air spaces
406 inside leaves; R_d is the respiration rate occurring during the day not related to photorespiration; Γ is
407 the CO_2 compensation point in the absence of R_d ; and J is the rate of photosynthetic electron
408 transport. These parameters are the same as those used in computing the CO_2 flux.

409 Our methods of computing stomatal and mesophyll conductances differ from previous studies
410 (*Suits et al.*, 2005; *Scholz et al.*, 2008; *Rayner et al.*, 2008) in the following ways: (1) these
411 conductances are calculated separately for sunlit and shaded leaves because BEPS is a two-leaf
412 model, in which the total GPP of a canopy is taken as the sum of sunlit and shaded leaf GPP; and (2)
413 the mesophyll conductance mechanistically depends on a set of parameters rather than being treated
414 as a constant or to be proportional to the stomatal conductance. Since it has been demonstrated that
415 sunlit and shaded leaf separation is essential for accurate modeling of canopy-level photosynthesis
416 (*Chen et al.*, 1999; *Sprintsin et al.*, 2011), it is expected that this separation is also essential for $^{13}\text{CO}_2$
417 flux estimation. We found that the use of Harley's method for computing the mesophyll conductance
418 makes the calculated ^{13}C photosynthetic fractionation stable for its global application, while the
419 simpler method of treating the mesophyll conductance in proportion with the stomatal conductance
420 often incurs abnormally large or small values of ^{13}C photosynthetic fractionation.

421 The photosynthetic $^{13}\text{CO}_2$ flux is in disequilibrium with the respiratory $^{13}\text{CO}_2$ flux because of
422 the change in atmospheric $^{13}\text{CO}_2$ concentration since the preindustrial time (*Ciais et al.*, 1995b; *Fung*
423 *et al.*, 1997). The heterotrophic respiratory flux from the decomposition of organic matter of different
424 ages carries the memory of the past atmospheric $^{13}\text{CO}_2$ concentration, while the photosynthetic $^{13}\text{CO}_2$
425 flux is affected by the current atmospheric $^{13}\text{CO}_2$ concentration. The isotopic composition of each of
426 the 9 soil carbon pools ($\delta^{13}C_{\text{soil}, i}$) is estimated with following formula:

$$427 \quad \delta^{13}C_{soil,i} = \delta^{13}C_a(2003-\tau_i) - \epsilon_{lph} \quad (16)$$

428 where $\delta^{13}C_a$ is the isotopic composition of carbon in atmosphere CO₂ in the past as determined by the
 429 ice-cord record (*Francey et al.*, 1999); ϵ_{lph} is the annual mean of photosynthetic discrimination in
 430 2003; and τ_i is the age of carbon pool i (Table 2) (*Ju et al.*, 2005). In the calculation of the mean age
 431 of a carbon pool, we have considered the ages of various carbon pools at the time of entering the pool
 432 (*Potter et al.*, 1993), so that the mean age is considerably larger than the turnover time determined by
 433 the decomposition rate (*Fung et al.*, 1997). The mean $\delta^{13}C_{soil}$ is taken as the flux-weighted $\delta^{13}C_{soil,i}$
 434 for the 9 carbon pools. The results of $\delta^{13}C_{soil}$ for the globe are shown in Figure 5. The ¹³C
 435 composition of the biosphere δ_{lb} in Eq. 8 is taken as the mean $\delta^{13}C_{soil}$, while the biospheric ¹³C
 436 composition δ_{lb}^e in equilibrium with the current atmosphere is taken as $\delta_a - \epsilon_{lph}$.

437 The accuracy of the BEPS model in simulating atmospheric ¹³CO₂ concentration was
 438 previously tested (*Chen et al.*, 2006; *Chen and Chen*, 2007) against measurements over a boreal
 439 forest at Fraserdale, Ontario, Canada (49°52'29.9''N, 81°34'12.3''W). Flask measurements of $\delta^{13}C_a$
 440 were made 40 times in both daytime and nighttime on a tower at a height of 20 m during a 3-day
 441 campaign on 21-23 July 1999. BEPS simulated these measurements with RMSE=0.34‰ and $r^2=0.76$.

442 **2.3 Transport modeling**

443 A transport-only version of the atmospheric chemistry and transport model TM5 (Krol et al.,
 444 2003; Krol et al., 2005) is used for CO₂ and ¹³CO₂ transport modeling to produce a fully linear
 445 operator on these fluxes. The spatial resolution of TM5 is 6°×4° for the globe and 3°×2° for North
 446 America, and the atmosphere is divided vertically into 25 layers with 5 layers in the planetary
 447 boundary layer. Tracer transport (advection, vertical diffusion, cloud convection) in TM5 is driven
 448 by offline meteorological fields taken from the European Centre for Medium Range Weather

449 Forecast (ECMWF) model. All physical parameterizations in TM5 are kept the same as the ECMWF
450 formulation to achieve compatibility between them. The four background fluxes from terrestrial
451 ecosystems, oceans, fossil-fuel burning, and biomass burning are individually inputted to TM5 to
452 calculate the contributions of these fluxes to the atmospheric CO₂ and ¹³CO₂ concentrations. Since
453 the main purpose of this study is to develop a joint inversion system, only one transport model is
454 used, the transport matrix **M** is assumed to be free of errors.

455 **2.4 CO₂ and ¹³CO₂ datasets**

456 Monthly CO₂ and ¹³CO₂ concentration data from 2000 to 2004 are compiled from the
457 GLOBALVIEW CO₂ and ¹³CO₂ database. Though the GLOBALVIEW database consists of both
458 extrapolated and interpolated data that were created based on the technique devised by *Masarie and*
459 *Tans* [1995], we selected the synchronized and smoothed values of actual observations to compile
460 our concentrations datasets. Only direct measurements of CO₂ from the GlobalView dataset are used
461 in our inversion after using a time-frequency weighting scheme (Deng and Chen, 2011). There are
462 5431 monthly data from 209 sites for 42 months used for CO₂ (5431 out of 8778, i.e. 209×42), and
463 3066 monthly data from 73 sites for ¹³CO₂ (i.e. 73×42 monthly data). Since the number of ¹³CO₂
464 observation sites is much smaller than that of CO₂ sites, all monthly data at 73 sites are used for
465 ¹³CO₂, and the missing ¹³CO₂ data are filled with the reference data provided in the same
466 GlobalView dataset. The filled data may have introduced an additional error to the dataset as shown
467 in Figure 15b.

468 To minimize the nonlinear aggregation effects of the large regions (*Pickett-Heaps, 2007*), the
469 contributions of the four background fluxes are subtracted from the above monthly concentrations.
470 So the matrix **c** in Eqs. (3) and (4) is expressed as

471
$$\mathbf{c} = \mathbf{c}_{\text{obs}} - \mathbf{c}_{\text{ff}} - \mathbf{c}_{\text{bio}} - \mathbf{c}_{\text{ocn}} - \mathbf{c}_{\text{fire}} \quad (17)$$

472 where \mathbf{c}_{obs} is the monthly CO₂ and ¹³CO₂ concentrations obtained from GLOBALVIEW, and \mathbf{c}_{ff} ,
473 \mathbf{c}_{bio} , \mathbf{c}_{ocn} , and \mathbf{c}_{fire} are simulated contributions of CO₂ and ¹³CO₂ concentrations from the terrestrial
474 biosphere, ocean, fossil-fuel, and fire fluxes, respectively.

475

476 3. Results

477 3.1 Prior CO₂ and ¹³CO₂ fluxes

478 Terrestrial ecosystem models integrate many sources of information, including vegetation
479 structure, soil, and meteorology, to estimate carbon exchange of the land surface with the atmosphere.
480 Prior CO₂ and ¹³CO₂ fluxes produced by a model can therefore provide indispensable constraints to
481 the otherwise ill-posed inversion based on CO₂ and ¹³CO₂ concentration observations alone.
482 Depending on the assigned relative magnitudes of the error matrixes of these observations and these
483 prior fluxes (i.e., \mathbf{R} and \mathbf{Q} in Eq. 3), these prior fluxes can have equal or even dominant importance
484 to these observations in the inversion results. We have therefore paid a great attention in modeling
485 these prior fluxes, in order to minimize the total inversion errors. Figure 2a shows an example of the
486 global terrestrial GPP distribution in 2003 modeled by BEPS. The total GPP in this year is 132±22
487 Pg C y⁻¹ (*Chen et al.*, 2012). This value is larger than some of the recent estimates, such as 123 Pg C
488 y⁻¹ by *Beer et al.* (2010), mostly because the LAI values used as input to BEPS are generally larger
489 than those of the MODIS product (*Garrigues et al.*, 2008). Our LAI values are larger because we
490 used a global clumping index map derived from a multi-angle satellite sensor POLDER (*Chen et al.*,
491 2005). Clumping increases shaded leaves which contributed about 35% to the total GPP globally.
492 Without considering this clumping effect, the shaded leaf area is underestimated, resulting in an

493 underestimation of the global GPP by 9% (*Chen et al.*, 2012). As the spatial distribution of clumping
494 is not uniform (boreal and tropical forests are most clumped and crops and grasses are least clumped),
495 this refinement in the GPP spatial distribution would have some effects on the inversion results
496 between regions.

497 The net ecosystem productivity (NEP), which is the difference between GPP and ecosystem
498 respiration modeled by BEPS, is shown in Figure 2b for 2003. Even though GPP has a large
499 uncertainty (globally 22 Pg C y⁻¹ by BEPS), the uncertainty in NEP is much smaller (globally 2 Pg C
500 y⁻¹ by BEPS) because a model spin-up approach is used to estimate the soil carbon pool sizes based
501 on a dynamic equilibrium assumption. Under this assumption, the annual heterotrophic respiration
502 (F_{lb}) equals annual NPP during the preindustrial period, and the soil carbon pool sizes are derived
503 from F_{lb} by solving a set of differential equations describing the decomposition and interactions
504 among the pools (*Govind et al.*, 2011). In this way, F_{lb} is forced to depend on NPP and the systematic
505 biases in GPP are not carried into NEP estimation. NEP is non-zero after the preindustrial period
506 because of the changes in climate and atmospheric composition (CO₂ and nitrogen) as well as
507 disturbance. In our regional modeling, both disturbance and non-disturbance effects are considered
508 for Canada (*Chen et al.*, 2003) and USA (*Zhang et al.*, 2012) forests. However, in our global model
509 spin-up from 1901 (taken as the end of preindustrial period) to 2000, only the non-disturbance effects
510 are considered because of lack of spatially explicit disturbance data outside of North America, while
511 carbon emission due to fire disturbance in the study period from 2000 to 2004 is considered
512 separately using the GFED dataset (*Randerson et al.*, 2007; *van der Werf et al.*, 2006). The prior net
513 CO₂ fluxes for the globe for the years 2002-2004 are given in Table 3 with inversion results with and
514 without the ¹³C constraint.

515 The global distribution of the total photosynthetic discrimination ($\delta^{13}C_{pt} = \delta^{13}C_a - \Delta$) modeled
516 by BEPS is shown in Figure 3. Forests, such as those in North America, Russia, Europe, Amazon,
517 central Africa, central China and southeast Asia, generally have high photosynthetic discrimination
518 rates (>16‰), while grassland and cropland (in particular C4 grasses and crops) have low
519 discrimination rates. Also shown in Figure 3 is the ocean diffusive discrimination against $^{13}CO_2$. The
520 discrimination over ocean is much smaller than that over land. This difference between land and
521 ocean discrimination may be considered as the largest signal of $^{13}CO_2$ observations on the global
522 carbon cycle (*Tans et al., 1990; Rayner et al., 2008*) and is considered in our inversion using
523 different $^{13}CO_2$ discrimination rates for ocean and land regions (see Eq. 6).

524 To estimate the disequilibrium between photosynthetic and respiratory discrimination against
525 $^{13}CO_2$, the global distribution of the mean soil carbon age is computed after weighting the ages of the
526 9 soil carbon pools against their fluxes due to decomposition (Figure 4). Forests at high latitudes
527 have the soil carbon age of about 40-60 years, while the tropical forests have much lower values in
528 the range from 10 to 30 years. This latitudinal distribution pattern is mostly determined by soil
529 temperature. In low latitudes, high temperature induces fast turnovers of detritus and fast soil carbon
530 pools, while at high latitudes, low temperature maintains relatively large fractions of slow and
531 passive soil carbon pools. Cropland and grassland also have larger fractions of fast and detritus
532 carbon pools than forest cover types and therefore have younger soil carbon on average. This spatial
533 distribution of soil carbon age has a strong influence on the total respiratory discrimination against
534 ^{13}C ($\delta^{13}C_r$) calculated by BEPS (Figure 5). Respiration from older carbon at high latitudes carries the
535 memory of the older atmosphere with less $^{13}CO_2$ concentration and hence has lower discrimination
536 rates (larger $\delta^{13}C_r$ or smaller absolute value). However, respiration would mostly depend on the

537 photosynthetic discrimination rates as soil organic matter originates from photosynthetic production.
538 As a result, forested areas have higher respiratory discrimination rates (lower $\delta^{13}\text{C}_r$, or larger absolute
539 value). Most of the high values of $\delta^{13}\text{C}_r$ in Figure 5 are associated with large fractions of C4 plants
540 in the grid, such as the corn belt in the USA, cropland in northeast China, southern border of Sahara
541 desert, and southeast South America. The global distribution of the disequilibrium between
542 photosynthetic and respiratory discrimination, taken as the difference between Figure 3 and Figure 5,
543 is shown in Figure 6. The disequilibrium is the largest at high latitude boreal forests in North
544 America and Eurasia because their soil carbon is the oldest, as shown in Figure 4. The spatial
545 distribution pattern of the disequilibrium is similar to those of *Ciais et al.* (1995b) and *Fung et al.*
546 (1997) but the magnitude is larger because the date of our result in 2000 is more recent than these
547 two previous studies. As the time lapses, the atmosphere is getting lighter in terms of the isotopic
548 composition of CO_2 resulting from the increased air-borne CO_2 from fossil fuel consumption. Also
549 shown in Figure 6 is the disequilibrium over the ocean estimated using the method of *Ciais et al.*
550 (1995b). This ocean disequilibrium has a large latitudinal gradient because of the gradients in sea
551 surface temperature gradient and the fluxes of CO_2 and $^{13}\text{CO}_2$. The spatial distribution in the
552 disequilibrium and the differences in disequilibrium between ocean and land may be considered to be
553 the secondary signal of $^{13}\text{CO}_2$ observations on the global carbon cycle. The effects of these
554 disequilibria on the carbon flux are considered in our inversion through presubtracting their
555 contributions to the measured $^{13}\text{CO}_2$ composition in Eq. 10.

556 **3.2 Inverse modeling results**

557 Although the inversions were made for the 2000-2004 period, the results of the first two years are not
558 included in the analysis because they are affected by the assumption of uniform global distributions

559 of CO₂ and ¹³CO₂ concentrations at the start of our transport modeling using TM5. An 18-24 month
560 period is usually considered to be necessary for the simulated distributions to reach realistic states
561 with reasonably accurate prior surface fluxes from ocean and land and atmospheric transport
562 simulations (*Rödenbeck et al.*, 2003; *Deng and Chen*, 2011). The following results are therefore
563 summarized as the average for the 2002-2004 period.

564 3.2.1 Partition between ocean and land sinks with and without ¹³CO₂ constraint

565 To investigate the usefulness of ¹³CO₂ observations in partitioning between ocean and land
566 sinks, we conducted inversions with and without ¹³CO₂ constraint as expressed in Eq. 6, i.e. with and
567 without the ¹³C-related expansions of the matrixes. The CO₂-only inversion increases the land sink
568 from the prior of 2.61 PgC y⁻¹ to 3.40 PgC y⁻¹ while decreasing the ocean sink from the prior of 2.13
569 PgC y⁻¹ to 1.48 PgC y⁻¹ (Table 3). These results are similar to those of *Deng and Chen* (2011). The
570 results from the joint inversion are considerably different: the posterior sinks for land and ocean
571 become 2.53 and 2.36 PgC y⁻¹ (Table 3), respectively, suggesting that the use of ¹³CO₂ observations
572 in the inversion considerably influenced the partition between land and ocean fluxes. The ratio
573 between land and ocean sinks is 1.07. The joint inversion system developed in this study may be
574 regarded as a different form of double deconvolution. Using the double deconvolution method with
575 the global average disequilibrium coefficients of 0.49‰ and 0.78‰ and the disequilibrium fluxes of
576 26.8 PgC y⁻¹ ‰ and 66 PgC y⁻¹ ‰ for land and ocean derived in this study (Table 4), respectively,
577 we also calculated the land and ocean sinks to be 2.90 and 2.36 PgC y⁻¹, respectively. The ratio
578 between land and ocean sinks is 1.23, which is close to the value of 1.07 derived from the joint
579 inversion system, indicating that the joint inversion can effectively perform double deconvolution.
580 Our joint inversion system differs from previous double deconvolution systems (*Siegenthaler and*
581 *Oeschger*, 1987; *Keeling et al.*, 1989a; *Francey et al.*, 1995; *Randerson et al.*, 2002) in the following

582 ways: (1) the estimation of CO₂ fluxes for the land and ocean is additionally constrained by the prior
583 fluxes for the land and ocean rather than entirely dependent on measured CO₂ concentration and
584 ¹³CO₂ composition; and (2) the spatio-temporal variations in all parameters associated with isotopic
585 discrimination and disequilibrium are considered in the estimation of the CO₂ flux using a
586 mechanistic biospheric model rather than global average values or simple models based on covariates.
587 These differences in methodology as well as the differences in the mean disequilibrium fluxes may
588 explain why the ocean and land sinks from the joint inversion system differ from the various double
589 deconvolution results.

590 The impacts of ¹³CO₂ data on the joint inversion can also be evaluated from the view point of
591 global ¹³CO₂ mass budget. Table 5 shows the budgets and its components for the prior, double
592 deconvolution, CO₂-only inversion and joint inversion cases. In these cases, the isofluxes due to
593 fossil fuel emission, land and ocean disequilibrium, and atmospheric storage change are the same,
594 and only those due to discrimination over land and ocean are adjusted. The prior case shows a global
595 imbalance of -5.0 PgC y⁻¹ ‰, indicating that either the prior land or ocean fluxes or both are
596 inconsistent with ¹³CO₂ measurements. Through double deconvolution, this imbalance is greatly
597 reduced to 0.8 PgC y⁻¹ ‰, mostly by an increase in the discrimination flux over land because of its
598 large discrimination rate. The CO₂-only inversion increases the land discrimination flux while
599 decreasing the ocean discrimination flux, resulting in no improvement in the global isotopic balance.
600 The joint inversion optimized both ocean and land fluxes in the direction consistent with ¹³CO₂
601 measurements, reducing the imbalance considerably to 1.8 PgC y⁻¹ ‰. These cases illustrate clearly
602 that the global isotopic mass balance is very sensitive to the partition between ocean and land fluxes
603 because of the large difference in the discrimination rate between land and ocean. In this analysis, the

604 disequilibrium fluxes are not adjusted, but the influences of the uncertainties in these fluxes on the
605 inversion results are analyzed in Section 3.2.4.

606 Existing estimates for the ocean sink for anthropogenic CO₂ in 2000's varies from 1.94 to 2.6
607 Pg C y⁻¹ (Wanninkhof *et al.*, 2013; Landchuster *et al.*, 2014; Majkut *et al.*, 2014; DeVries, 2014). The
608 average ocean sink for the 2002-2004 period summarized by the Global Carbon Project (GCP) (Le
609 Quéré *et al.*, 2013) is 2.4 Pg C y⁻¹, while the land sink in the same period is 2.7 Pg C y⁻¹ as the
610 residual of the global carbon budget after including the emission due to land use change as a source
611 of carbon. Although the prior estimates of these sinks in our inversions are similar to these values,
612 our CO₂-only inversion considerably increases the land sink and decreases the ocean sink. The
613 addition of ¹³CO₂ measurements to the inversion significantly decreases the land sink and increases
614 the ocean sink, pulling the inversion results in the direction to agree with these existing estimates
615 (Figure 7). This may indicate that the use of ¹³CO₂ measurements in the joint inversion has improved
616 the CO₂ estimation. In this comparison, we have not considered the unknown small amount (0.1-0.3
617 Pg C y⁻¹) of lateral carbon transport in rivers from land to ocean. This amount is included in some of
618 the estimates of the ocean sink used by GCP, and therefore should be subtracted from the ocean sink
619 and added to the land sink by GCP in order to compare with our atmospheric inversion results.

620 3.2.2 Influence of ¹³CO₂ constraint on the spatial distribution of the inverted carbon flux

621 The ¹³CO₂ constraint not only modified the partition between ocean and land fluxes but also
622 their spatial distribution patterns. Figure 8 shows the result of the CO₂-only inversion (i.e. without
623 the ¹³CO₂ constraint), as the net carbon flux over land and ocean averaged for the period of 2002-
624 2004. Figure 9 shows the difference between inversions with and without the ¹³CO₂ constraint, i.e.
625 the result of CO₂+¹³CO₂ inversion minus that of CO₂-only inversion. The general patterns of the

626 inverted carbon flux are similar between these two inversions because these inversions depend
627 primarily on the CO₂ concentration, the prior flux, the error matrixes of the prior flux, and
628 concentration observations. However, there are several large or notable differences: (1) The Amazon
629 region (Region 31) is changed from a carbon source to a sink (Figure 10. Note: a reduction in sources
630 is shown as a negative value); (2) the carbon sink in the tropical Asia (Region 37) is noticeably
631 reduced (by about 10-20 gC m⁻²y⁻¹ from a sink magnitude of about 80-100 gC m⁻²y⁻¹); (3) The sink in
632 Asia (Region 36) decreases pronouncedly by about 10-20 gC m⁻²y⁻¹, while the sinks in Russia
633 (Region 35) and Europe (Region 39) are also reduced by some extents (about 5-20 gC m⁻²y⁻¹); (4)
634 most small regions in the southern part of North America show increases in sinks, but those in the
635 northern part (Canada and Alaska) show increases in sources (see also Figure 11). The overall sink
636 in North America decreases from 0.67 to 0.54 Pg C y⁻¹ (Figure 10); and (5) most ocean regions at
637 mid-latitudes have small gains in sink.

638 It is of particular importance to note that the ¹³CO₂ constraint changed the Amazon region
639 from a carbon source of 0.43 ± 0.46 Pg C y⁻¹ to a carbon sink of 0.08 ± 0.38 Pg C y⁻¹ with a notable
640 reduction in the posterior uncertainty, which is higher than uncertainty reductions in most other
641 regions (Figure 10). This change is likely caused by the relatively large addition of information from
642 ¹³CO₂ in this tropical region where CO₂ observations are sparse, causing large uncertainties in the
643 inverted flux in this region in the CO₂-only inversion. *Potter et al.* (2009) simulated the net
644 ecosystem productivity (NEP) of the Amazon region using the CASA model driven by remote
645 sensing inputs and found that the NEP for the region was slightly negative (-0.07 Pg C y⁻¹) over the
646 2000-2004 period. *Davidson et al.* (2012) summarized from various inventory-based studies that
647 mature forests in the region was accumulating carbon at a rate of 0.29-0.57 Pg C y⁻¹ over the decade
648 before 2005, meaning that NEP is positive. Since the fire emission is estimated to be 0.50 Pg C y⁻¹

649 (*Richey et al.*, 2002), the Amazon region would be either net source of carbon or about carbon
650 neutral. Since spatially explicit fire emission is considered together with fossil fuel emission as a
651 source in our study, the inverted carbon flux corresponds to $-NEP$, and therefore the result from our
652 joint inversion is in broad agreement with Potter's and Davidson's results. Without the $^{13}CO_2$
653 constraint, our inversion result shows an unreasonably large source of carbon in the Amazon region.

654 3.2.3 Influence of the spatial distribution of photosynthetic discrimination on the inverted carbon 655 flux

656 The joint inversion results shown in Figures 9 to 11 are from Case I with the best estimates of
657 the ^{13}C discrimination and disequilibrium fluxes and therefore represent a baseline study to which
658 other cases are compared for the purpose of investigating the importance of accurate consideration of
659 the spatial distributions of isotopic discrimination and disequilibrium over land and ocean. Case II is
660 designed to investigate the importance of considering the spatial distribution of the photosynthetic
661 isotopic discrimination over land for inverting the CO_2 flux by fixing the discrimination at a constant
662 over land. Figure 12a shows the spatial distribution of the difference in the total isotopic
663 discrimination, i.e. $D_j = \varepsilon_{iph,j}$, among 39 land regions between Case I and Case II, calculated as Case
664 I minus Case II. Regions with positive differences in D_j are shown with positive differences in the
665 inverted CO_2 flux (Figure 12b), meaning larger sinks (negative values) in Case II, and vice versa.
666 This is because a smaller discrimination rate (smaller than -14.1%) means a larger CO_2 flux from the
667 atmosphere to the surface (more negative value) for the same change in $^{13}CO_2$ concentration in the
668 atmosphere. Under the same condition, a larger discrimination induces a smaller sink (less negative).
669 The absolute regional differences between Case I and Case II are considerable (Figure 12b), e.g. up
670 to $18 \text{ g C m}^{-2}\text{y}^{-1}$, showing increases in sinks in Africa, Asia and Australia and decreases in sinks in

671 Amazon, Europe, Russia and most of the small regions in North America. However, the total global
672 sink values of Case II after ignoring the spatial distribution of the disequilibrium rate over land
673 change very little from those of Case I (Table 3): from 2.53 ± 0.93 to 2.49 ± 0.95 Pg C y^{-1} for land and
674 from 2.36 ± 0.49 to 2.35 ± 0.48 Pg C y^{-1} for ocean. This is because the global mean discrimination rates
675 are the same between these two cases.

676 3.2.4 Influence of the uncertainties in disequilibrium fluxes on the inverted carbon flux

677 The average disequilibrium coefficients and fluxes for land and ocean derived in this study are
678 comparable to published results (Table 4), although the estimates of the disequilibrium flux over
679 ocean in previous studies vary in a large range. The uncertainty in the estimated land and ocean
680 disequilibrium fluxes mainly arises from two sources: the estimated disequilibrium coefficient and
681 one-way CO₂ flux from the surface. Mathematically, the total uncertainty in the disequilibrium flux,
682 denoted as $\Delta(\delta \cdot F)$, equals $\sqrt{(\Delta\delta \cdot F)^2 + (\delta \cdot \Delta F)^2}$. For land, the first source depends on the modeled
683 mean soil carbon age by BEPS, which is estimated to be ± 5 years, causing an error in the
684 disequilibrium coefficient to be $\pm 0.11\%$ based on the slope of δ_a against time at about 1979 (the
685 flux-weighted global mean soil carbon age is 24 years). The second source is estimated to be 9.5 PgC
686 y^{-1} in NPP, which is taken as 45% of the error in GPP, i.e. 21 PgC y^{-1} (Chen *et al.*, 2012). With
687 NPP=59.4 PgC y^{-1} and the mean disequilibrium coefficient of 0.49‰, the uncertainty in the estimated
688 land disequilibrium flux is therefore $\sqrt{(0.11 \times 59.4)^2 + (0.49 \times 9.5)^2} = 8.0$ PgC y^{-1} ‰. For ocean, the
689 error in the modeled disequilibrium coefficient is mostly caused by sea surface temperature (SST), if
690 the coefficients in the equation developed by *Ciais et al.* (1995b) are assumed to be accurate. With an
691 error of 1.0K in SST, the error in the calculated global average disequilibrium coefficient is $\pm 0.12\%$.
692 The error in one-way the ocean flux is difficult to estimate, but we use the value of 10 PgC y^{-1}

693 inferred from the global isotopic budget uncertainty by *Alden et al. (2010)*. Their inferred range of
694 the ocean disequilibrium flux is from 92.3 to 100.2 PgC y⁻¹ ‰, and we use our disequilibrium
695 coefficient of 0.78‰ to calculate this one-way flux uncertainty. Based on the OPA-PISCES-T model,
696 the one-way flux from ocean to atmosphere is 84.6 PgC y⁻¹, and the uncertainty in the estimated
697 ocean disequilibrium flux is therefore $\sqrt{(0.12 \times 84.6)^2 + (0.78 \times 10)^2} = 12.7 \text{ PgC y}^{-1} \text{ ‰}$.

698 Case III, Case IV and Case V are conducted to investigate the relative importance of the
699 disequilibrium fluxes over land and ocean (Table 3) in the CO₂ flux inversion. In Case III, where the
700 disequilibrium over land is ignored while other settings remain the same as Case I, the land sink
701 increases by 1.05 PgC y⁻¹, while the ocean sink decreases by 0.08 PgC y⁻¹ in comparison with Case I.
702 When the disequilibrium over ocean is ignored instead (Case IV), the land sink increases by 0.13
703 PgC y⁻¹, while the ocean sink increases by 2.08 PgC y⁻¹, in comparison with Case I. When the
704 disequilibria over both land and ocean are ignored, the land sink increases by 1.18 PgC y⁻¹, while the
705 ocean sink increases by 1.96 PgC y⁻¹, in comparison with Case I. Results from these case studies
706 suggest that in the joint inversion using both CO₂ and ¹³CO₂ measurements, the inverted CO₂ flux
707 can be significantly influenced by the disequilibrium fluxes of land and ocean. The carbon sinks over
708 land and ocean increase when these disequilibrium fluxes are ignored because the photosynthetic and
709 diffusive sources of ¹³CO₂ have to increase to make up for the shortfall due to ignoring the
710 disequilibrium sources. These pronounced influences of the disequilibrium fluxes on the CO₂ sink
711 inversion suggest that ¹³CO₂ data contain strong signals for the global carbon cycle. In the joint
712 inversion, these data can have the power to distort the global CO₂ mass balance if the ¹³CO₂ mass
713 budget (Eq. 8) is not properly simulated. The influence of ¹³CO₂ on the joint inversion depends only
714 weakly on the estimated uncertainty in the ¹³CO₂ data. We found that if the uncertainty is reduced by

715 half, the sum of the land and ocean sink deviates from the CO₂-only case by 2-6% for all scenarios,
716 suggesting that the mean disequilibrium fluxes play the dominant roles in the joint inversion.

717 The impacts of these disequilibrium fluxes on the inverted CO₂ flux determined in Case III,
718 Case IV and Case V are similar to previous results using the double deconvolution technique (*Tans et*
719 *al.*, 1993; *Ciais et al.*, 1995b; *Randerson et al.*, 2002). However, the influences of these
720 disequilibrium fluxes on the joint inversion could possibly be compromised due to the small number
721 of ¹³C observation sites relative to the number of CO₂ observation sites used in the joint inversion.
722 The number of linear equations for CO₂ concentration in our joint inversion system (Eq. 6) greatly
723 exceeds the number for ¹³C composition, with a potential of dampening the impact of ¹³C data on the
724 inverted results. To investigate the possibility of this dampening effect, we conducted a set of
725 inversions using ¹³C data alone (Table 6) and found that the impacts of the disequilibrium fluxes on
726 the inversion results are similar to those of the joint inversion. In Case V shown in Table 6, for
727 example, ignoring the disequilibrium fluxes causes the land sink to increase by 1.06 PgC y⁻¹ and
728 ocean sink to increase by 2.37 PgC y⁻¹, resulting in a total increase of 3.43 PgC y⁻¹, which is similar
729 to the total difference of 3.14 PgC y⁻¹ produced by the joint inversion. These similar results suggest
730 that ¹³C data used in the way described by Eqs. 6-8 have played the expected role in the joint
731 inversion. By comparing results shown in Tables 3 and 6, it is also encouraging to see that inversions
732 using ¹³C data alone can produce reasonable results for the CO₂ flux, although we believe that the
733 joint inversion results shown in Table 3 are more reliable. Our finding on the usefulness of the small
734 ¹³CO₂ dataset somewhat confirms the claim of Enting et al. (1993 and 1995) that the temporal trend
735 in ¹³CO₂ concentration is the major signal constraining the partition between ocean and land sinks.

736 According to the difference of the inverted flux between Case III to Case I, the uncertainty of
737 8.0 PgC y⁻¹ % in the land disequilibrium flux would cause an uncertainty of 0.47 PgC y⁻¹ in the land

738 flux. According to the comparison between Case IV to Case I, the uncertainty of $12.7 \text{ PgC y}^{-1} \%$ in
739 the ocean disequilibrium flux would cause an uncertainty of 0.54 PgC y^{-1} in the ocean flux. These
740 uncertainties in the land and ocean fluxes are 17% and 24% of the jointly inverted fluxes for land and
741 ocean (Case I in Table 3), respectively. The impact of the uncertainty in the disequilibrium flux over
742 land is only slightly smaller than the posterior uncertainty of the inverted land flux, but the impact
743 over ocean is larger than the posterior uncertainty.

744 **Discussion**

745 After the CO_2 fluxes are optimized through the inversions, the posterior CO_2 concentration at
746 all stations in each month can be calculated from Eq. 2, and similarly the posterior $^{13}\text{CO}_2$
747 composition can also be calculated from Eq. 10 by replacing the prior discrimination fluxes with
748 posterior discrimination fluxes. One way to evaluate the effectiveness of the joint inversion is to
749 examine the improvement in the posterior CO_2 and $^{13}\text{CO}_2$ concentrations against measurements.
750 Figure 13 shows concentrations for 10 randomly selected stations from different regions, which are
751 indicated in Figure 1. The CO_2 and $^{13}\text{CO}_2$ concentrations produced using the prior fluxes
752 considerably deviate from observations at all stations. The posterior CO_2 concentration from the
753 CO_2 -only inversion shows great improvements over the prior concentration in comparison with
754 observations. The posterior CO_2 concentration from the joint inversion does not differ significantly
755 from that of the CO_2 -only inversion. At some stations the joint inversion produces slightly lower root
756 mean square differences (RMSD) against observations, but in some stations the opposite is true, as
757 indicated by the RMSD values shown in the header of each plot. It is expected that in some stations,
758 the posterior CO_2 concentration in the joint inversion can be slightly worsened because of the
759 influence of $^{13}\text{CO}_2$. The posterior $^{13}\text{CO}_2$ concentration is pronouncedly improved over the prior in
760 comparison with observations and almost mimics the observed magnitudes and temporal variations,

761 indicating that the joint inversion system can forcefully adjust CO₂ fluxes to match with ¹³CO₂
762 observation through the prescribed discrimination rates. The posterior CO₂ concentrations for either
763 CO₂-only or joint inversion show larger seasonal amplitudes than observations at northern
764 hemisphere stations, although the means are about the same as observations. This suggests that both
765 carbon uptake during the growing season and ecosystem respiration in the non-growing season might
766 have been overestimated, even though the annual net carbon flux may be unbiased. Further work is
767 needed to constrain the large photosynthetic and respiratory fluxes separately rather than the net flux
768 only.

769 In order to provide a comprehensive evaluation, the posterior CO₂ and ¹³CO₂ concentrations at
770 all stations are shown in Figures 14 and 15 against observations. In Figure 14, we see pronounced
771 improvements in the posterior concentrations from both the CO₂-only and joint inversions over the
772 prior case. However, the improvements of these two inversions are similar (the joint inversion has a
773 smaller intercept and a slope closer to one, but the CO₂-only inversion has a slightly larger r² value).
774 This is in agreement with the cases shown for the individual stations: some stations are improved and
775 some worsened by the use of ¹³CO₂ data, manifesting the force of this additional data on the
776 inversion. In Figure 15, the posterior ¹³CO₂ concentration from the joint inversion is shown to be
777 greatly improved from the prior case. In the joint inversion, the increase of the posterior land and
778 ocean sinks over the prior sinks that remove CO₂ from the atmosphere logically corrects for the
779 positive bias in the CO₂ concentration produced using the prior fluxes (Figure 14a). The posterior
780 concentration correlation with observation is stronger for ¹³CO₂ than for CO₂, indicating that
781 isofluxes are effectively optimized in the joint inversion according to ¹³CO₂ data. However, some
782 points in Figure 15b scatter greatly from the 1:1 line, and these points are mostly likely the missing

783 data filled with the reference data (Section 2.4). As other error sources cannot be excluded, these data
784 are retained in our inversion.

785 After adding $^{13}\text{CO}_2$ data to the inversion system, the uncertainty in the inverted CO_2 flux
786 increased from 0.84 to 0.93 PgC y^{-1} for land and from 0.40 to 0.49 PgC y^{-1} for ocean (Table 3,
787 difference between the CO_2 -only case and Case I), i.e. 11% and 23% increases in uncertainty for land
788 and ocean, respectively. The relative error in preprocessed $^{13}\text{CO}_2$ measurements used in the joint
789 inversion is considerably larger than that in CO_2 measurements, causing these increases in the
790 uncertainty of jointly inverted CO_2 fluxes from the CO_2 -only case. The $^{13}\text{CO}_2$ measurements were
791 preprocessed before the inversion as the remaining concentration after removing the contributions of
792 fossil fuel emission and prior land and ocean discrimination and disequilibrium fluxes (Eq. 10), and
793 therefore they contain uncertainties from these contributions in addition to measurement
794 uncertainties. Errors in modeling the spatial and temporal variations of the $^{13}\text{CO}_2$ flux stem from
795 many sources including errors in modeling the discrimination, which is affected by the fractionation
796 of the $^{13}\text{CO}_2$ flow through leaf boundary layer, stomata, mesophyll, etc., and the disequilibrium,
797 which depends on the sizes of 9 soil carbon pools and their ages. Although the ocean $^{13}\text{CO}_2$
798 discrimination is small, its disequilibrium has a strong latitudinal gradient, which is approximately
799 calculated using the mean monthly temperature. The error in the calculated ocean disequilibrium
800 coefficient is estimated to be $\pm 1.2\%$ for the monthly values at a given location and $\pm 0.12\%$ for the
801 global annual total. Because of these errors, we estimate that the relative uncertainty in the prior
802 $^{13}\text{CO}_2$ fluxes is similar to that of the prior CO_2 flux over both land and ocean.

803

804 4. Conclusion

805 The usefulness of atmospheric $^{13}\text{CO}_2$ measurements at 73 stations for global carbon cycle
806 estimation is explored through their use as an additional constraint on an atmospheric inversion of
807 the surface carbon flux using CO_2 observations. The following conclusions are drawn from this
808 study:

- 809 1. This ^{13}C constraint on the joint inversion considerably alters the partition between land and
810 ocean sinks obtained from CO_2 -only inversion, decreasing the land sink from 3.40 ± 0.84 to
811 2.53 ± 0.93 Pg C y^{-1} , while increasing the ocean sink from 1.48 ± 0.40 to 2.36 ± 0.49 Pg C y^{-1} for
812 the 2002-2004 period. Over land, this alteration induces the largest sink increases in the
813 Amazon region and largest source increases in southern Africa and Asia, where CO_2
814 observations are sparse and therefore the additional signal from $^{13}\text{CO}_2$ data becomes most
815 important. Over ocean, sink increases are found broadly at middle and high latitudes in both
816 hemispheres.
- 817 2. The spatial distribution of the $^{13}\text{CO}_2$ discrimination rate over land has considerable impacts
818 on the spatial distribution of the inverted CO_2 sink over land (up to 15% in some regions),
819 suggesting that reliable models for simulating the spatial distribution of the ^{13}C discrimination
820 rate over land are needed for effective use of $^{13}\text{CO}_2$ data for global carbon cycle inversion.
- 821 3. The joint inversion is sensitive to the $^{13}\text{CO}_2$ disequilibrium fluxes over both land and ocean.
822 Ignoring these fluxes in the joint inversion causes the inverted total land and ocean sink to
823 increase by 1.18 and 1.96 PgC y^{-1} , respectively. The uncertainty in our disequilibrium flux
824 calculation is estimated to be 8.0 $\text{PgC y}^{-1}\%$ and 12.7 $\text{Pg C y}^{-1}\%$ for land and ocean,
825 respectively, inducing an uncertainty in the inverted flux of 0.47 Pg C y^{-1} for land and 0.54 Pg
826 C y^{-1} for ocean.

827 **Acknowledgement**

828 We greatly appreciate the GlobalView dataset that is available from the NOAA Climate
829 Monitoring and Diagnostics Laboratory. NCEP/DOE 2 Reanalysis data are provided by the
830 NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at
831 <http://www.esrl.noaa.gov/psd/>. This research was support by a grant (2010CB950704) from the
832 Global Change Key Program of the Chinese Ministry of Science and Technology and a grant (GR-
833 646) from the former Canadian Foundation for Climate and Atmospheric Sciences. Drs. Misa
834 Ishizawa, Douglas Chan and Kaz Higuchi of Environment Canada (EC) provided useful suggestions
835 and technical assistance at the beginning of the project, and Jonathan Fung assisted in the CO₂-only
836 inversion during his M.Sc. studies. We are also indebted to Drs. Pieter Tans of NOAA and Lin
837 Huang of EC for their guidance in the development and validation of the isotope module in BEPS in
838 the early years. Dr. Peter Rayner was a critical and constructive reviewer of several versions of this
839 paper and provided high-level intellectual inputs to this study which are greatly appreciated.

840 **4. References**

- 841 Alden, C., J. B. Miller, and J. W. C. White (2010), Can bottom-up ocean CO₂ fluxes can be
842 reconciled with atmospheric 13C observations?, *Tellus B*, 62B, 369-388.
843
844 Andres R.J., G. Marland, T. Boden T., and S. Bischof (2000), Carbon dioxide emissions from fossil
845 fuel consumption and cement manufacture, 1751-1991, and an estimate of their isotopic composition
846 and latitudinal distribution, in Wigley T.M.L., Schimel D.S. (eds.) *The Carbon Cycle*. Cambridge:
847 Cambridge University Press. 53-62.
848
849 Aumont, O., E. Maier-Reimer, S. Blain, and P. Monfray (2003), An ecosystem model of the global
850 ocean including Fe, Si, P colimitations, *Global Biogeochem. Cycles*, 17(2), 1060.
851
852 Baker, D. F., R. M. Law, K. R. Gurney, P. Rayner, P. Peylin, A. D. Denning, P. Bousquet, L.
853 Bruhwiler, Y. H. Chen, P. Ciais, I. Y. Fung, M. Heimann, J. John, T. Maki, S. Maksyutov, K.
854 Masarie, M. Prather, B. Pak, S. Taguchi, and Z. Zhu (2006), TransCom 3 inversion intercomparison:
855 Impact of transport model errors on the interannual variability of regional
856 CO₂ fluxes, 1988–2003, *Global Biogeochem. Cy.*, 20, GB1002, doi:10.1029/2004GB002439.

857
858 Ball, J. T. (1988), An analysis of stomatal conductance, Ph.D. thesis, 89 pp., Stanford Univ.,
859 Stanford, Calif.

860 Beer, C., M. Reichstein, E. Tomelleri, P. Ciais, M. Jung, N. Carvalhais, C. Rodenbeck, M.A. Arain,
861 D. Baldocchi, G.B. Bonan, A. Bondeau, A. Cescatti, G. Lasslop, A. Lindroth, M. Lomas, S.
862 Luysaert, H. Margolis, K.W. Oleson, O. Roupsard, E. Veenendaal, N. Viovy, C. Williams, F.I.
863 Woodward, and D. Papale (2010), Terrestrial Gross Carbon Dioxide Uptake: Global Distribution and
864 Covariation with Climate, *Science* 329:834-838.

865 Buitenhuis, E., C. Le Quéré, O. Aumont, G. Beaugrand, A. Bunker, A. Hirst, T. Ikeda, T. O'Brien, S.
866 Piontkovski, and D. Straile (2006), Biogeochemical fluxes through mesozooplankton, *Global*
867 *Biogeochem. Cycles*, 20(2), GB2003.

868 Canadell, J. G., C. Le Quéré, M. R. Raupach, C. B. Field, E. T. Buitenhuis, P. Ciais, T. J. Conway,
869 N. P. Gillett, R. A. Houghton, and G. Marland (2007), Contributions to accelerating atmospheric CO₂
870 growth from economic activity, carbon intensity, and efficiency of natural sinks, *Proceedings of the*
871 *National Academy of Sciences*, 104(47), 18866-18870.

872
873 Chen, B., J. M. Chen, P. Tabs and L. Huang (2006), Simulating dynamics of ¹³CO₂ in the planetary
874 boundary layer over a boreal forest region: An approach to estimate carbon isotope rectification.
875 *Tellus* 58B, 537-549.

876
877 Chen, B., and J. M. Chen, 2007. Interannual Temporal variation in carbon isotope discrimination in
878 response to meteorological and physiological driving factors in a boreal forest ecosystem, *Plant, Cell*
879 *and Environment*, 30: doi: 10.1111/j.1365-3040.2007.01707.

880 Chen, J. M., J. Liu, J. Cihlar, and M. L. Goulden (1999), Daily canopy photosynthesis model through
881 temporal and spatial scaling for remote sensing applications, *Ecological Modelling*, 124(2-3), 99-119.
882

883 Chen, J. M., W. Ju, J. Cihlar, D. Price, J. Liu, W. Chen, J. Pan, T. A. Black, and A. Barr (2003),
884 Spatial distribution of carbon sources and sinks in Canada's forests based on remote sensing, *Tellus B*.
885 55(2): 622-642.

886 Chen, J. M., C. H. Menges, and S. G. Leblanc (2005), Global derivation of the vegetation clumping
887 index from multi-angular satellite data, *Remote Sensing of Environment*, 97: 447-457.

888 Chen, J. M., G. Mo, J. Pisek, J. Liu, F. Deng, M. Ishizawa, and D. Chan (2012), Effects of foliage
889 clumping on the estimation of global terrestrial gross primary productivity, *Global Biogeochem.*
890 *Cycles*, 26, GB1019, doi:10.1029/2010GB003996.

891 Chen, J. M., J. Fung, G. Mo, F. Deng, and T. West (2015), Atmospheric inversion of the global
892 surface carbon flux with consideration of the spatial distribution of US crop production and
893 consumption. *Biogeosciences*, 12, 323-343.

894

895 Ciais, P., P. P. Tans, M. Trolier, J.W. C. White, and R. J. Francey (1995a). A large northern-
896 hemisphere terrestrial CO₂ sink indicated by the ¹³C/¹²C ratio of atmospheric CO₂. *Science*, 269,
897 1098-1102.
898

899 Ciais, P., P. P. Tans, J. W. C. White, M. Trolier, R. J. Francey, J. A. Berry, D. R. Randall, P. J.
900 Sellers, J. G. Collatz, and D. S. Schimel (1995b), Partitioning of ocean and land uptake of CO₂ as
901 inferred by $\delta^{13}\text{C}$ measurements from the NOAA Climate Monitoring and Diagnostics Laboratory
902 Global Air Sampling Network, *J. Geophys. Res.*, 100(D3), 5051–5070, doi:10.1029/94JD02847.
903

904 Davidson, E. A., A. C. de Araujo, P. Artaxo et al., (2012), The Amazon basin in transition. *Nature*,
905 481, 321-328.
906

907 Deng, F., J. M. Chen, M. Ishizawa, C.-W. Yuen, G. Mo, K. A. Z. Higuchi, D. Chan, and S.
908 Maksyutov (2007), Global monthly CO₂ flux inversion with a focus over North America, *Tellus B*,
909 59(2), 179-190.
910

911 Deng, F., J. M. Chen, S. Plummer, M. Chen, and J. Pisek (2006), Algorithm for global leaf area
912 index retrieval using satellite imagery, *IEEE Transactions on Geoscience and Remote Sensing*, 44(8),
913 2219-2229.
914

915 Deng, F., and J. M. Chen (2011), Recent global CO₂ flux inferred from atmospheric CO₂
916 observations and its regional analysis. *Biogeosciences*, 8: 3263–3281.

917 DeVries, T. (2014), The oceanic anthropogenic CO₂ sink: Storage, air-sea fluxes, and transports over
918 the industrial era, *Global Biogeochem. Cycles*, 28, 631–647, doi:10.1002/2013GB004739.
919

920 Enting, I.G., C.M. Trudinger, R.J. Francey and H. Granek (1993), Synthesis inversion of atmospheric
921 CO₂ using the GISS tracer transport model, *Division of Atmospheric Research technical paper*,
922 No.29, CSIRO Australia.
923

924 Enting, I. G., C. M. Trudinger, and R. J. Francey (1995), A synthesis inversion of the concentration
925 and $\delta^{13}\text{C}$ of atmospheric CO₂. *Tellus B*, 47: 35–52.
926

927 Enting, I. G. (2002), *Inverse Problems in Atmospheric Constituents transport*, Cambridge University
928 Press.
929

930 Farquhar, G. D., S. von Caemmerer, and J. A. Berry (1980), A biochemical model of photosynthetic
931 CO₂ assimilation in leaves of C3 species, *Planta*, 149, 78–90, doi:10.1007/BF00386231.
932

933 Farquhar, G. D., and R. A. Richards (1984), Isotopic composition of plant carbon correlates with
934 water-use efficiency of wheat genotypes, *Aust., J. Plant. Physiol.*, 11, 539–552.

935 Farquhar G.D., J. R. Ehleringer and K. T. Hubick (1989), Carbon isotope discrimination and
936 photosynthesis. *Annual Review of Plant Physiology and Plant Molecular Biology*. 40:503–537.

937 Francey, R. J., P. P. Tans, C. E. Allison, I. G. Enting, J. W. C. White, and M. Troler (1995),
938 Changes in oceanic and terrestrial carbon uptake since 1982. *Nature*, 373, 326–330.
939

940 Francey R. J., C. E. Allison, D. M. Etheridge, C. M. Trudinger, I. G. Enting, M. Leuenberger, R. L.
941 Langenfelds, E. Michel, L. P. Steele (1999), A 1000-year high precision record of $\delta^{13}\text{C}$ in
942 atmospheric CO_2 , *Tellus B*, 51(2), 170–193.
943

944 Fung, I.Y., C. B. Field, J. A. Berry, M. V. Thompson, J. T. Randerson, C. M. Malmström, P. M.
945 Vitousek, G. James Collatz, P. J. Sellers, D. A. Randall, A. S. Denning, F. Badeck (1997), Carbon-
946 ^{13}C exchanges between the atmosphere and biosphere, *Global Biogeochemical Cycles*, 11: 507-533.

947 Garrigues S., R. Lacaze, F. Baret, J.T. Morisette, M. Weiss, J.E. Nickeson, R. Fernandes, S.
948 Plummer, N.V. Shabanov, R.B. Myneni, Y. Knyazikhin, W. Yang (2008), Validation and
949 intercomparison of global Leaf Area Index products derived from remote sensing data, *J. Geophys.*
950 *Res.*, 113, G02028:doi:10.1029/2007JG000635.

951 GLOBALVIEW-CO2C13: Cooperative Atmospheric Data Integration Project - $\delta^{13}\text{C}$ of Carbon
952 Dioxide. CD-ROM, NOAA ESRL, Boulder, Colorado [Also available on Internet via anonymous
953 FTP to ftp.cmdl.noaa.gov, Path: ccg/co2c13/GLOBALVIEW], 2009.

954 Govind, A., and J. M. Chen (2011), Spatially distributed modeling of the long-term carbon balance
955 of a boreal landscape. *Ecological Modeling*, doi:10.1016/j.ecolmodel.2011.04.007.

956 Gurney, K. R., R. M. Law, A. S. Denning, P. J. Rayner, D. Baker, P. Bousquet, L. Bruhwiler, Y. H.
957 Chen, P. Ciais, S.M. Fan, I. Y. Fung, Manuel Gloor, Martin Heimann, K Higuchi, J. John, T Maki, S.
958 Maksyutov, K. Masarie, P. Peylin, M. Prather, B. C. Pak, J. T. Randerson, J. L. Sarmiento, S.
959 Taguchi, T Takahashi, C. W. Yuen (2002), Towards robust regional estimates of CO_2 sources and
960 sinks using atmospheric transport models, *Nature*, 415(6872), 626-630.

961 Gurney, K. R., K. R. Gurney, R. M. Law, A. S. Denning, P. J. Rayner, D. Baker, P. Bousquet, L.
962 Bruhwiler, Y. H. Chen, P. Ciais, S.M. Fan, I. Y. Fung, M. Gloor, M. Heimann, K. Higuchi, J. John, E.
963 Kowalczyk, T. Maki, S. Maksyutov, P. Peylin, M. Prather, B. C. Pak, J. L. Sarmiento, S. Taguchi, T.
964 Takahashi, C. W. Yuen (2003), TransCom3 CO_2 inversion intercomparison: 1. Annual mean control
965 results and sensitivity to transport and prior flux information, *Tellus B*, 55(2), 555-579.
966

967 Harley, P. C., F. Loreto, G. Di Marco, and T. D. Sharkey (1992), Theoretical Considerations when
968 Estimating the Mesophyll Conductance to CO_2 Flux by Analysis of the Response of Photosynthesis
969 to CO_2 , *Plant Physiol.* 98:1429–1436.

970 Houghton, R.A. (2007), Balancing the global carbon budget, *Annual Review of Earth and Planetary*
971 *Sciences* 35:313-347.
972

973 Ise, T., C. M. Litton, C. P. Giardina, and A. Ito (2010), Comparison of modeling approaches for
974 carbon partitioning: impact on estimates of global net primary production and equilibrium biomass of

975 woody vegetation from MODIS GPP. *Journal of Geophysical Research*, Vol., 115, G04025,
976 doi:10.1029/2010JG001326.

977 Jacobson, A., S. Fletcher, N. Gruber, J. Sarmiento, and M. Gloor (2007), A joint atmosphere-ocean
978 inversion for surface fluxes of carbon dioxide: 2. Regional results, *Global Biogeochem. Cycles*, 21(1),
979 GB1020, doi:10.1029/2006GB002703.

980

981 Ju, W., and J. M. Chen (2005), Distribution of soil carbon stocks in Canada's forests and wetland
982 simulated based on drainage class, topography and remote sensing. *Hydrological Processes*, 19:77-
983 94.

984

985 Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G.
986 White, J. Woollen, Y. Zhu, A. Leetmaa, B. Reynolds, M. Chelliah, W. Ebisuzaki, W. Higgins, J.
987 Janowiak, K. C. Mo, C. Ropelewski, J. Wang, R. Jenne, and D. Joseph (1996), The NCEP/NCAR
988 40-year reanalysis project, *Bull. Amer. Meteor. Soc.*, 77, 437-471.

989

990 Kanamitsu, M., W. Ebisuzaki, J. Woollen, S-K Yang, J. J. Hnilo, M. Fiorino, and G. L. Potter.
991 (2002), NCEP-DEO AMIP-II Reanalysis (R-2), *Bulletin of the American Meteorological Society*,
992 1631-1643.

993

994 Keeling, C. D., R. B. Bacastow, A. F. Carter, S. C. Piper, T. P. Whorf, and co-authors (1989a). A
995 three-dimensional model of atmospheric CO₂ transport based on observed winds: 1. analysis of
996 observational data. In: Aspects of climate variability in the Pacific and Western Americas (ed.
997 Peterson, D. H.). American Geophysical Union, Washington, D.C., 165-236.

998

999 Keeling, C. D., S. C. Piper, and M. Heimann (1989b). A three-dimensional model of atmospheric
1000 CO₂ transport based on observed winds: 4. Mean annual gradients and interannual variations. In:
1001 Aspects of climate variability in the Pacific and Western Americas (ed. Peterson, D. H.). American
1002 Geophysical Union, Washington D.C., 305-363.

1003

1004 Krol, M. C., J. Lelieveld, D. E. Oram, G. A. Sturrock, S. A. Penkett, C. A. M. Brenninkmeijer, V.
1005 Gros, J. Williams, and H. A. Scheeren (2003), Continuing emissions of methyl chloroform from
1006 Europe, *Nature*, 421(6919), 131-135.

1007 Krol, M. C., S. Houweling, B. Bregman, M. van den Broek, A. Segers, P. van Velthoven, W. Peters,
1008 F. Dentener, and P. Bergamaschi (2005), The two-way nested global chemistry-transport zoom
1009 model TM5: algorithm and applications, *Atmos. Chem. Phys.*, 5(2), 417-432.

1010 Landschützer, P., N. Gruber, D. C. E. Bakker, and U. Schuster, (2014), Recent variability of the
1011 global ocean carbon sink. *Global Biogeochemical Cycles*, 28, 927–949, doi:10.1002/2014GB004853.

1012 Le Quéré, C., Andres, R. J., Boden, T., Conway, T., Houghton, R. A., House, J. I., Marland, G.,
1013 Peters, G. P., van der Werf G. R., Ahlstrom, A., Andrew, R. M., Bopp, L., Canadell, J. G., Ciais, P.,
1014 Doney, S. C., Enright, C., Friedlingstein, P., Huntingford, C., Jain, A. K., Jourdain, C., Kato, E.,
1015 Keeling, R. F., Klein G. K., Levis, S., Levy, P., Lomas, M., Poulter, B., Raupach, M. R., Schwinger,

1016 J., Sitch, S., Stocker, B. D., Viovy, N., Zaehle, S., and Zeng, N. (2013). [The global carbon budget](#)
1017 [1959–2011](#). *Earth Syst. Sci. Data* 5: 165–185, doi:10.5194/essd-5-165-2013.

1018 Liu, J., J. M. Chen, J. Cihlar, and W. M. Park (1997), A process-based boreal ecosystem productivity
1019 simulator using remote sensing inputs, *Remote Sensing of Environment*, 62(2), 158-175.
1020

1021 Madec, G., P. Delecluse, M. Imbard, and C. Lévy (1998), OPA 8.1 ocean general circulation model
1022 reference manual, *Notes du pôle de modélisation IPSL*, 91pp, <http://www.lodyc.jussieu.fr/opa/>.
1023

1024 Majkut, J. D., J. L. Sarmiento, and K. B. Rodgers (2014), A growing oceanic carbon uptake: results
1025 from an inversion study of surface pCO₂ data. *Global Biogeochemical Cycles*, 28, 335–351,
1026 doi:10.1002/2013GB004585.

1027

1028 Marland, G., T. A. Boden, and R. J. Andres (2009), Global, Regional, and National Fossil Fuel CO₂
1029 Emissions. In Trends: A Compendium of Data on Global Change. Carbon Dioxide Information
1030 Analysis Center, Oak Ridge National Laboratory, U.S. Department of Energy, Oak Ridge, Tenn.,
1031 U.S.A.

1032

1033 Masarie, K. A., and P. P. Tans (1995), Extension and integration of atmospheric carbon dioxide data
1034 into a globally consistent measurement record, *J. Geophys. Res.*, 100(D6), 11593-11610.
1035

1036 Michalak, A. M., A. Hirsch, L. Bruhwiler, K. R. Gurney, W. Peters, and P. P. Tans (2005),
1037 Maximum likelihood estimation of covariance parameters for Bayesian atmospheric trace gas surface
1038 flux inversions, *J. Geophys. Res.*, 110, D24107, doi:10.1029/2005JD005970.
1039

1040 Olivier, J. G. J., J. A. Van Aardenne, F. J. Dentener, V. Pagliari, L. N. Ganzeveld, and J. A. H. W.
1041 Peters (2005), Recent trends in global greenhouse gas emissions: regional trends 1970–2000 and
1042 spatial distribution of key sources in 2000, *Environmental Sciences*, 2(2), 81 - 99.
1043

1044 Park, G. H., R. Wanninkhof, S. C. Doney, T. Takahashi, K. Lee, R. A. Feely, C. L. Sabine, J.
1045 Trinanes, and I. D. Lima, (2010), Variability of global net sea-air CO₂ fluxes over the last three
1046 decades. *Chemical and Physical Meteorology*, 62, 352–368.
1047

1048 Patra, P. K., S. Maksyutov, M. Ishizawa, T. Nakazawa, T. Takahashi, and J. Ukita (2005),
1049 Interannual and decadal changes in the sea-air CO₂ flux from atmospheric CO₂ inverse modeling,
1050 *Global Biogeochem. Cycles*, 19(4), GB4013.

1051 Peters, W., J. B. Miller, J. Whitaker, S. A. Denning, A. Hirsch, M. Krol, D. Zupanski, L. Bruhwiler,
1052 and P. P. Tans (2005), An ensemble data assimilation system to estimate CO₂ surface fluxes from
1053 atmospheric trace gas observations, *J. Geophys. Res.*, 110,
1054 D24304, doi:10.1029/2005JD006157.

1055 Peters, W. A. R. Jacobson, C. Sweeney, A. E. Andrews, T.J. Conway, K. Masarie, J. B. Miller, L. M.
1056 P. Bruhwiler, G. Petron, A. I. Hirsch, D. E. J. Worthy, G.R Werf, J. T. Randerson, P. O. Wennberg,
1057 M.C. Krol, P. P. Tans (2007), An atmospheric perspective on North American carbon dioxide
1058 exchange: CarbonTracker, *Proceedings of the National Academy of Sciences*, 104(48), 18925-18930.

1059 Peylin, P., P. Bousquet, C. Le Quéré, S. Sitch, P. Friedlingstein, G. McKinley, N. Gruber, P. Rayner,
1060 and P. Ciais (2005), Multiple constraints on regional CO₂ flux variations over land and oceans.
1061 *Global Biogeochemical Cycles* 19: doi: 10.1029/2003GB002214.
1062

1063 Pickett-Heaps, C. A. (2007), Atmospheric CO₂ inversion cross-validation using non-surface CO₂
1064 data., *PhD dissertation*.

1065 Potter, C.S., J.T. Randerson, C.B., Field, P.A. Matson, P.M. Vitousek, H.A. Mooney, and S.A.
1066 Klooster (1993), Terrestrial ecosystem production: A process model based on global satellite and
1067 surface data, *Global Biogeochemical Cycles*, 7, 811-841.

1068 Potter, C., S. Klooster, A. Huete, V. Genovese, M. Bustamante, L. Guimaraes Ferreira, R. C. de
1069 Oliveira Jr., and R. Zepp, (2009), Terrestrial carbon sinks in the Brazilian Amazon and Cerrado
1070 region predicted from MODIS satellite data and ecosystem modeling. *Biogeosciences*, 6, 937–945.
1071

1072 Randerson, J. T., G. J. Collatz, J. E. Fessenden, A. D. Munoz, C. J. Still, J. A. Berry, I. Y. Fung, N.
1073 Suits, and A. S. Denning (2002), A possible global covariance between terrestrial gross primary
1074 production and ¹³C discrimination: Consequences for the atmospheric ¹³C budget and its response to
1075 ENSO, *Global Biogeochem. Cycles*, 16(4), 1136, doi:10.1029/2001GB001845
1076

1077 Randerson, J. T., G. R. van der Werf, L. Giglio, G. J. Collatz, and P. S. Kasibhatla (2007), Global
1078 Fire Emissions Database, Version 2 (GFEDv2.1). Data set. , Available on-line [<http://daac.ornl.gov/>]
1079 from Oak Ridge National Laboratory Distributed Active Archive Center, Oak Ridge, Tennessee,
1080 U.S.A.
1081

1082 Rayner, P. J, 2001. Atmospheric perspectives on the ocean carbon cycle. In E. D. Schulze, S. P.
1083 Harrison, M. Heimann, E. A. Holland, J. Lloyd, I. C. Prentice, and D. Schimel, editors, *Global*
1084 *biogeochemical cycles in the climate system*, pages 285–294. Academic Press, San Diego.

1085 Rayner, P. J., I. G. Enting, R. J. Francey, and R. L. Langenfelds (1999), Reconstructing the recent
1086 carbon cycle from atmospheric CO₂, δ¹³C and O₂/N₂ observations, *Tellus, Ser. B*, 51, 213– 232.
1087

1088 Rayner, P. J., R. M. Law, C. E. Allison, R. J. Francey, C. M. Trudinger, and C. Pickett-Heaps (2008),
1089 Interannual variability of the global carbon cycle (1992–2005) inferred by inversion of atmospheric
1090 CO₂ and δ¹³CO₂ measurements, *Global Biogeochem. Cycles*, 22, GB3008,
1091 doi:10.1029/2007GB003068.

1092 Reynolds, R. W and T. M. Smith (1994), Improved global sea surface temperature analyses using
1093 optimum interpolation. *J. Climate*, 7, 929–948.

1094 Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes and W. Wang (2002), An improved in situ
1095 and satellite SST analysis for climate. *J. Climate*, 15, 1609-1625.
1096

1097 Richey, J. E., Melack, J. M., Aufdenkampe, A. K., Ballester, V. M. & Hess, L. L. (2002), Outgassing
1098 from Amazonian rivers and wetlands as a large tropical source of atmospheric CO₂. *Nature* 416,
1099 617–620.
1100

1101 Rödenbeck, C., S. Houweling, M. Gloor, and M. Heimann (2003), CO₂ flux history 1982-2001
1102 inferred from atmospheric data using a global inversion of atmospheric transport, *Atmos. Chem.*
1103 *Phys.*, 3(6), 1919-1964.

1104 Rödenbeck, C., D. C. E. Bakker, N. Metzl, A. Olsen, C. Sabine, N. Cassar, F. Reum, R. F. Keeling,
1105 and M. Heimann, (2014), Interannual sea–air CO₂ flux variability from an observation-driven
1106 ocean mixed-layer scheme, *Biogeosciences Discuss.*, 11, 3167– 3207, doi:10.5194/bgd-11-3167-
1107 2014.
1108

1109 Scholze, M., P. Ciais, and M. Heimann (2008), Modeling terrestrial ¹³C cycling: Climate, land use
1110 and fire, *Global Biogeochem. Cycles*, 22, GB1009, doi:10.1029/2006GB002899.

1111 Siegenthaler, U. and H. Oeschger (1987). Biospheric CO₂ emissions during the past 200 years
1112 reconstructed by deconvolution of ice core data. *Tellus* 39B, 140-154.

1113 Sprintsin, M., J. M. Chen, and P. Czurylowicz (2011), Combining land surface temperature and
1114 shortwave infrared reflectance for early detection of mountain pine beetle infestations in western
1115 Canada. *Journal of Applied Remote Sensing* 5(1), 053566.

1116 Steinkamp, K., and N. Gruber (2013), A joint atmosphere-ocean inversion for the estimation of
1117 seasonal carbon sources and sinks, *Global Biogeochem. Cycles*, 27, doi:10.1002/gbc.20064.
1118

1119 Stephens, B., et al. (2007), Weak northern and strong tropical land carbon uptake from vertical
1120 profiles of atmospheric CO₂, *Science*, 316(5832), 1732–1735.
1121

1122 Suits, N. S., A. S. Denning, J. A. Berry, C. J. Still, J. Kaduk, J. B. Miller, and I. T. Baker (2005),
1123 Simulation of carbon isotope discrimination of the terrestrial biosphere, *Global Biogeochem. Cycles*,
1124 19, GB1017, doi:10.1029/2003GB002141.
1125

1126 Still, C. J., J. A. Berry, G. J. Collatz, and R. S. DeFries (2003), Global distribution of C₃ and C₄
1127 vegetation: Carbon cycle implications, *Global Biogeochem. Cycles*, 17(1), 1006, doi: 10.1029 / 2001
1128 GB001807.
1129

1130 Tans, P. P., I. Y. Fung, and T. Takahashi (1990), Observational Constrains on the Global Atmospheric
1131 CO₂ Budget, *Science*, 247(4949), 1431-1438.
1132

1133 Tans, P. P., J. A. Berry, and R. F. Keeling (1993), Oceanic ¹³C data: A new window on CO₂ uptake
1134 by the oceans, *Global Biogeochem. Cycles*, 7, 353-368.
1135

1136 Tarantola, A. (1987), *Inverse Problem Theory*, 605 pp., Elsevier, Amsterdam, The Netherlands.
1137

1138 van der Velde, I. R., J. B. Miller, K. Schaefer, K. A. Masarie, S. Denning, J. W. C. White, P. P. Tans,
1139 M. C. Krol, and W. Peters (2013), Biosphere model simulations of interannual variability in
1140 terrestrial $^{13}\text{C}/^{12}\text{C}$ exchange, *Global Biogeochemical Cycles*, 27, 637–649

1141 van der Werf, G. R., J. T. Randerson, L. Giglio, G. J. Collatz, P. S. Kasibhatla, and A. F. Arellano Jr.
1142 (2006), Interannual variability of global biomass burning emissions from 1997 to 2004, *Atmos. Chem.*
1143 *Phys. Discuss.*, 6(2), 3175-3226.

1144 Webb, R. S., C. E. Rosenzweig, and E. R. Levine (1991), A global data set of soil particle size
1145 properties, *NASA Tech. Memo.*, TM-4286, 40 pp.

1147 Wanninkhof, R., G. -H. Park, T. Takahashi, C. Sweeney, R. Feely, Y. Nojiri, N. Gruber, S. C. Doney,
1148 G. A. McKinley, A. Lenton, C. Le Quéré, C. Heinze, J. Schwinger, H. Graven, and S. Khatiwala,
1149 (2013), Global ocean carbon uptake: magnitude, variability and trends,
1150 *Biogeosciences*, 10, 1983-2000, doi:10.5194/bg-10-1983-2013,

1151 Zhang F., J. M. Chen, J. Chen, C. M. Gough, T. A. Martin, D. Dragoni (2012), Evaluating spatial and
1152 temporal patterns of MODIS GPP over the conterminous U.S. against flux measurements and a
1153 process model, *Remote Sensing of Environment*, Vol. 124: 717-729.

1154
1155 Zhang S.P., X. Yi, X. G. Zheng, Z. Q. Chen, et al., 2014. Global Carbon Assimilation System using a
1156 Local Ensemble Kalman Filter with Multiple Ecosystem Models, *Journal of Geophysical Research-*
1157 *Biogeosciences*, doi: 10.1002/2014JG002792.

1158
1159
1160
1161
1162

1163 **Table 1.** Biophysical parameters are assigned by plant functional types in BEPS. References for the
 1164 chosen values of these parameters are found in *Chen et al. (2012)*.

Parameters ^a	Broadleaf Evergreen	Broadleaf Deciduous	Evergreen Conifers	Deciduous Conifers	Shrub	C4 Plants	Others
$V_{\text{cmax}} \mu\text{mol m}^{-2} \text{s}^{-1}$ (at 25°C)	29.0±7.7	57.7±21.2	62.5±24.7	39.1±11.7	57.9±19.6	100.7±36.6	90.0±89.5
$J_{\text{max}} \mu\text{mol m}^{-2} \text{s}^{-1}$	55.1	123.7	135.2	79.2	124.1	193.1	200.0
$N \text{ g m}^{-2}$	2.17±0.8	1.74±0.71	3.10±1.35	1.81±0.64	1.86±0.84	1.62±0.61	1.69±0.69
$\chi_n \text{ m}^2 \text{ g}^{-1}$	0.48	0.59	0.33	0.56	0.57	0.62	0.60
Slope (m)	8	8	8	8	8	4	8
Intercept (b), $\text{mol m}^{-2} \text{s}^{-1}$	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011
LAI	4.07±2.02	3.14±1.99	3.05±1.62	2.42±1.45	1.49±1.06	1.55±1.22	1.64±1.15
Clumping Index	0.66±0.045	0.70±0.047	0.74±0.057	0.78±0.051	0.75±0.059	0.75±0.050	0.76±0.059
Canopy height (m)	23	23	20	20	4	4	4

1165 Where V_{cmax} is the leaf maximum carboxylation rate at 25°C, J_{max} is the maximum electron transport rate, N is the leaf nitrogen content, χ_n is the slope
 1166 of V_{cmax} variation with N , and m and b are the slope and intercept in the Ball-Berry equation. The peak growing season LAI and clumping index are
 1167 given as the mean and standard deviation for each plant functional type.
 1168

1169

1170

1171 **Table 2.** Global average ages of soil carbon pools computed by BEPS with consideration of the
 1172 influences of temperature and soil moisture on the decomposition rates of these pools.
 1173

1173

1174

Soil carbon pool i	Name	Global Average Age τ_i (yr)
1	Surface structural leaf litter	5.0
2	Surface metabolic leaf litter	2.3
3	Soil structural litter	4.4
4	Soil metabolic litter	2.3
5	Woody litter	34.9
6	Surface microbe	11.1
7	Soil microbe	28.5
8	Slow carbon	35.5
9	Passive carbon	667.9

1175

1176

1177 Table 3. Inverted fluxes (Pg C y^{-1}), averaged for 2002 –2004, for land and ocean regions with ($\text{CO}_2 +$
 1178 $^{13}\text{CO}_2$) and without (CO_2 only) ^{13}C constraint. The negative sign denotes the flux from the
 1179 atmosphere to the surface (sink). Various treatments are made to ^{13}C discrimination and
 1180 disequilibrium fluxes represented by the following cases:

1181 Case I: Full consideration of the regional differences in discrimination and disequilibrium;

1182 Case II: Same as Case I, but the annual photosynthetic discrimination ratio is set at a constant of -
 1183 14.1‰, although it's monthly variation pattern as modeled by BEPS is retained;

1184 Case III: Same as Case I, but the disequilibrium flux over land is ignored;

1185 Case IV: Same as Case I, but the disequilibrium flux over ocean is ignored;

1186 Case V: Same as Case I, but the disequilibrium flux over both land and ocean is ignored.

1187

1188

Region	Prior flux	Double De-convolution	Inverted CO_2 flux					
			CO_2 data	$\text{CO}_2 + ^{13}\text{CO}_2$ data				
				Case I	Case II	Case III	Case IV	Case V
Land	-2.61	-2.90	-3.40	-2.53	-2.49	-3.58	-2.66	-3.71
	± 2.07		± 0.84	± 0.93	± 0.95	± 0.93	± 0.93	± 0.93
Ocean	-2.13	-2.36	-1.48	-2.36	-2.35	-2.24	-4.44	-4.32
	± 0.67		± 0.40	± 0.49	± 0.48	± 0.49	± 0.49	± 0.49

1189

1190

1191

1192 Table 4. Comparison of land and ocean disequilibrium coefficients and disequilibrium fluxes
 1193 calculated in this study with those in previous studies.

1194

1195

Studies	Year	Land Disequilibrium Coefficient (‰)	Land Disequilibrium Flux (PgC y ⁻¹ ‰)	Ocean Disequilibrium Coefficient (‰)	Ocean Disequilibrium Flux (PgC y ⁻¹ ‰)
This study	2002-2004	0.49	26.8	0.78	66
Fung et al. (1997)	1988	0.33	N/A	N/A	N/A
Randerson et al. (2002)	1981-1994	0.33	20	0.6	55
Alden et al. (2010)	1991-2007	0.45-0.61	22.7-30.6	N/A	92.3-100.2 (globe total)
Van der Velde et al. (2013)	1991-2007	0.486	25.4	N/A	48.7
Francey et al. (1995)	1987	0.43	25.8	0.48	43.8

1196 Table 5. Global isotopic mass budgets averaged for the 2002-2004 period for the prior, double de-
 1197 convolution, CO₂-only inversion, and joint inversion (unit: Pg C y⁻¹ ‰). Also shown are ocean and
 1198 land net fluxes (unit Pg C y⁻¹) for these cases for comparison purposes. For the prior fluxes, the
 1199 component of each flux are indicated in the brackets. The isotopic coefficients are same among the
 1200 cases.

Isotopic terms	Prior	Double de-convolution	CO ₂ –only inversion	Joint inversion
$-C_a d(\delta_a)/dt$	15.0 [750 Pg C × (-0.02‰ y ⁻¹)]	15.0	15.0	15.0
$F_f (\delta_f - \delta_a)^*$	-153.7 [8.9 Pg C y ⁻¹ × (-17.27‰)]	-153.7	-153.7	-153.7
$-(F_{lph} - F_{lb})\epsilon_{lh}$	36.7 [2.6 Pg C y ⁻¹ × (-14.10‰)]	40.9	47.9	39.5
$F_{lb}(\delta_{lb} - \delta_{be})$	26.8 [54.7 Pg C y ⁻¹ × (-0.49‰)]	26.8	26.8	26.8
$-(F_{ao} - F_{oa})\epsilon_{ao}$	4.2 [2.1 Pg C y ⁻¹ × (-2.00‰)]	4.8	3.0	4.6
$F_{oa} (\delta_{oa}^e - \delta_{oa})$	66.0 [84.6 Pg C y ⁻¹ × (-0.78‰)]	66.0	66.0	66.0
Global Budget	-5.0	0.8	-5.0	1.8
$(F_{lph} - F_{lb}), Pg C y^{-1}$	-2.6	-2.9	-3.4	-2.8
$(F_{ao} - F_{oa}), Pg C y^{-1}$	-2.1	-2.4	-1.5	-2.3

1201 * F_f is the carbon emission from fossil fuel and biomass burning, 6.9 and 2.1, Pg C y⁻¹, respectively,
 1202 and δ_f is weighted average ¹³C composition for fossil fuel and biomass burning, being 25.27‰, and δ_a = -8.0‰.

1203

1204
 1205 Table 6. Inverted fluxes (Pg C y^{-1}), averaged for 2002 –2004, for land and ocean regions using ^{13}C
 1206 data only. The negative sign denotes the flux from the atmosphere to the surface (sink). Various
 1207 treatments are made to ^{13}C discrimination and disequilibrium fluxes represented by the cases outlined
 1208 in Table 3.

1209

Region	Prior flux	Inverted CO_2 flux				
		$^{13}\text{CO}_2$ data				
		Case I	Case II	Case III	Case IV	Case V
Land	-2.61	-2.60	-2.56	-3.61	-2.65	-3.66
	± 2.07	± 0.96	± 0.99	± 0.96	± 0.96	± 0.96
Ocean	-2.13	-2.28	-2.27	-2.28	-4.65	-4.65
	± 0.67	± 0.53	± 0.54	± 0.53	± 0.53	± 0.53

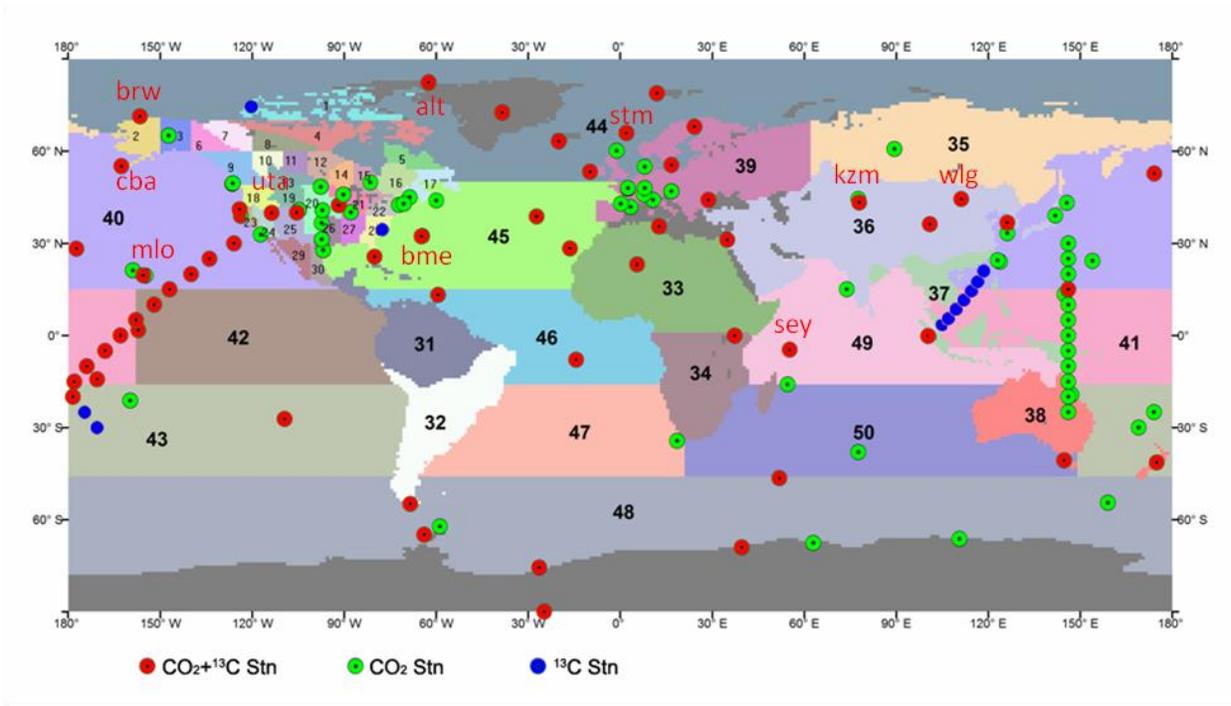
1210

1211

1212

1213

1214 **Figures**



1215

1216 Figure 1. A global nested inversion system with a focus in North America, in which oceans are
1217 divided into 11 regions and land areas are divided into 9 large and 30 small regions outside and
1218 within North America, respectively. Also shown are CO₂ and ¹³CO₂ observation stations included in
1219 the GlobalView database and used in this study. 10 of the stations are marked with their names
1220 because they are selected to compare prior and posterior concentrations in Figure 11.

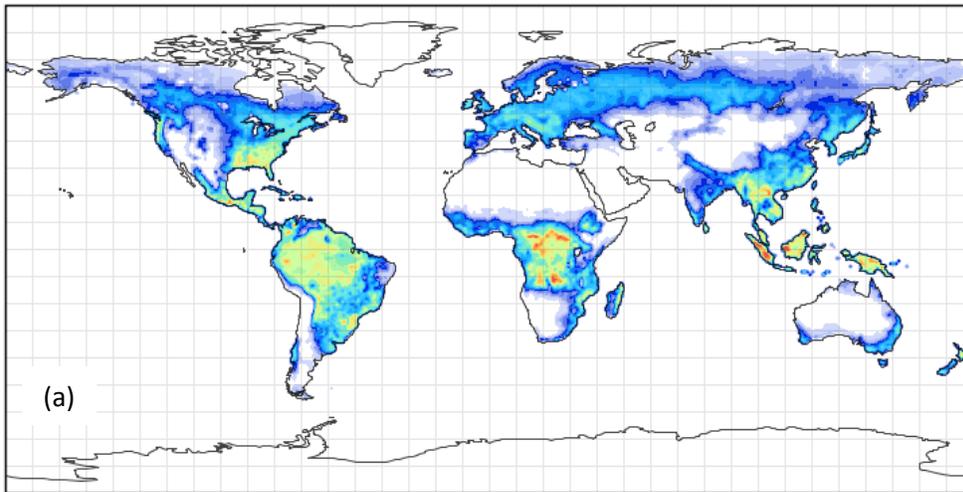
1221

1222

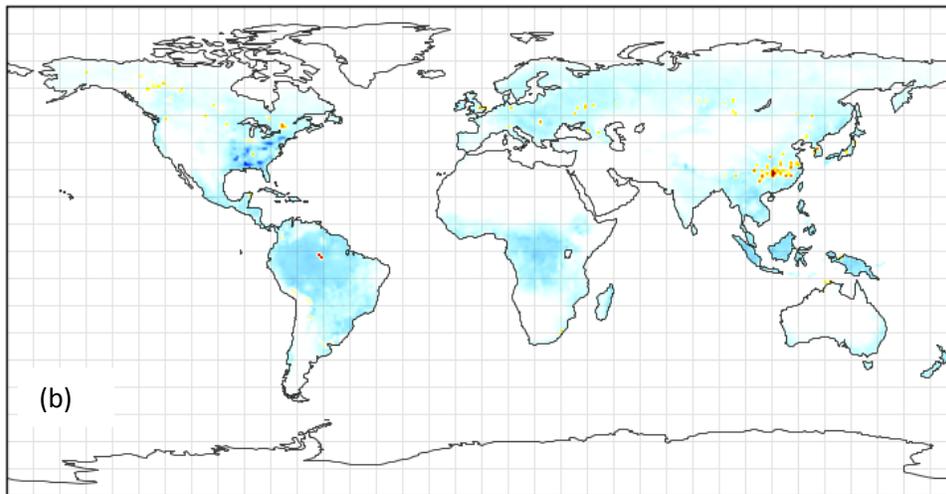
1223

1224

1225

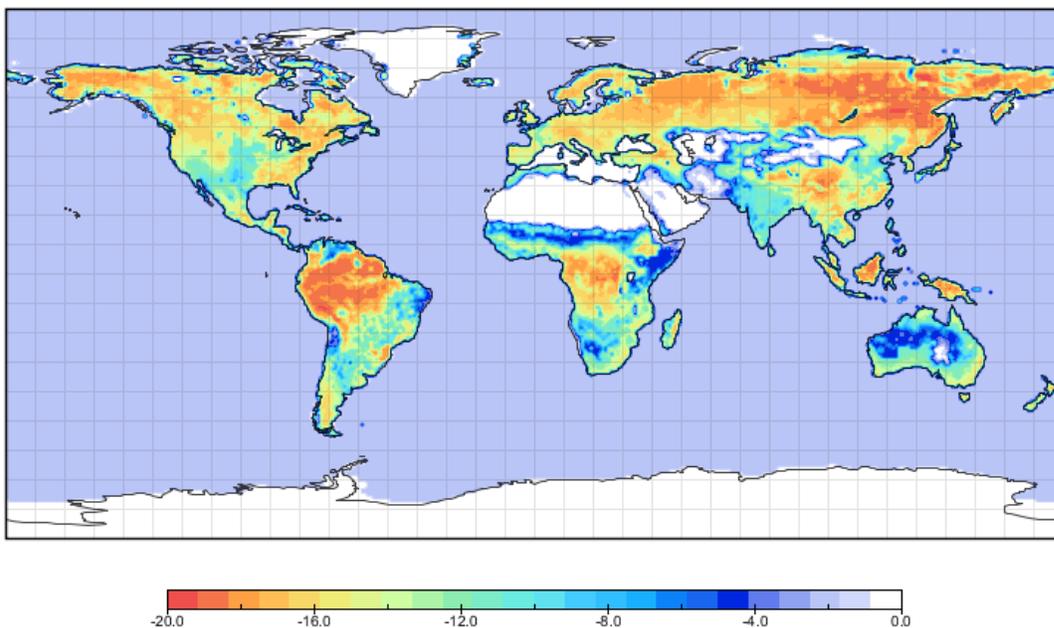


1226



1227

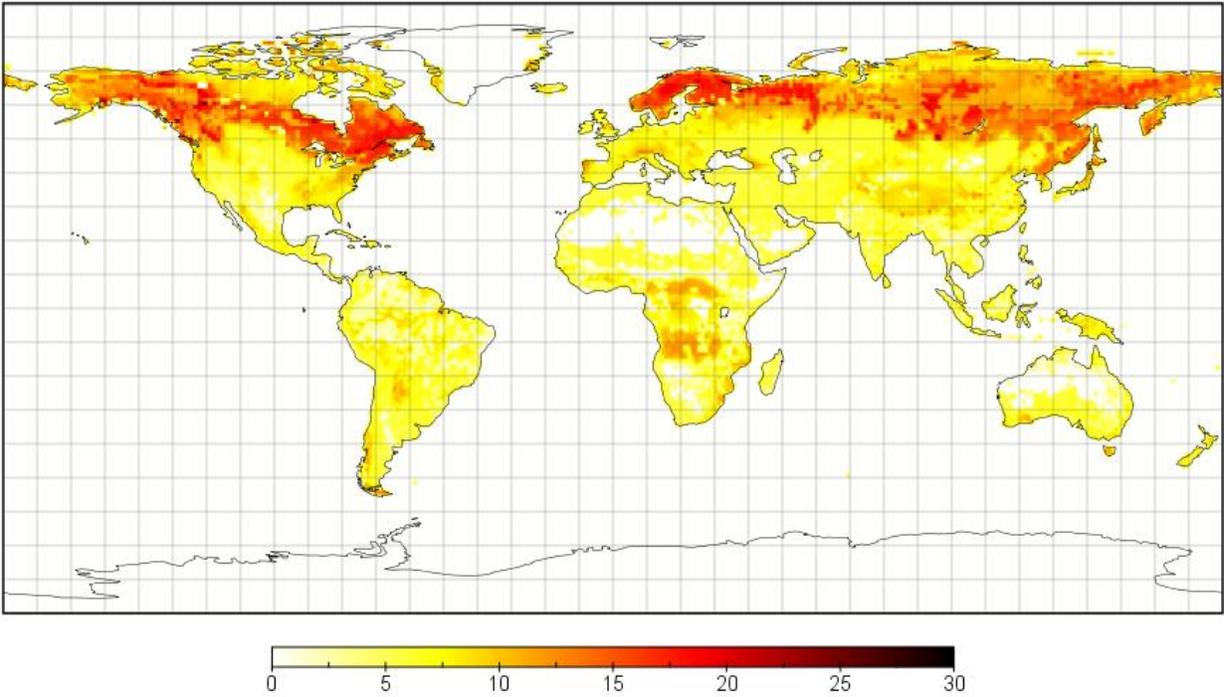
1228 Figure 2. (a) gross primary productivity (GPP) distribution in 2003 computed using remote sensing
1229 LAI and land cover maps and climate and soil data, and (b) net ecosystem productivity (NEP)
1230 distribution in 2003. Both are calculated using the BEPS model. Annual NEP maps from 2000 to
1231 2004 are used to as the prior flux in the inversions. This GPP map is used to distribute the flux
1232 uncertainty among the 39 land regions.



1233

1234 Figure 3. The annual mean of the total photosynthetic ^{13}C discrimination (Δ in Eq. 7) in 2003.

1235



1236

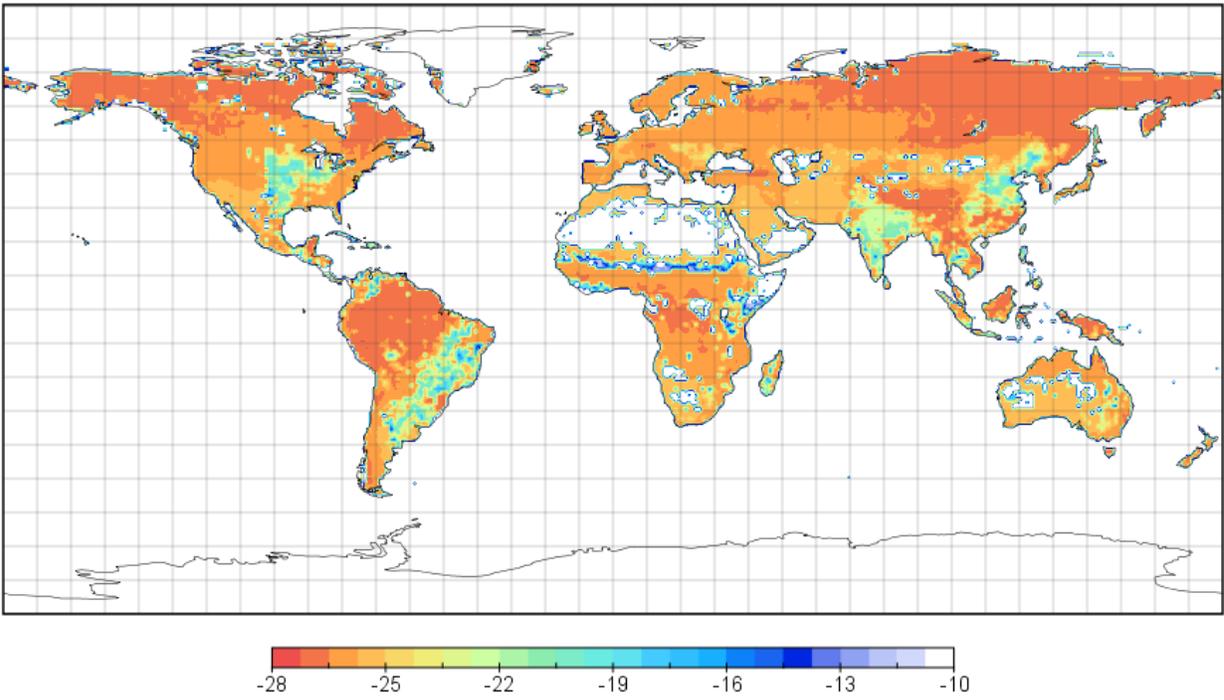
1237 Figure 4. Global distribution of the flux-weighted mean age of soil carbon pools (Eq. 8).

1238

1239

1240

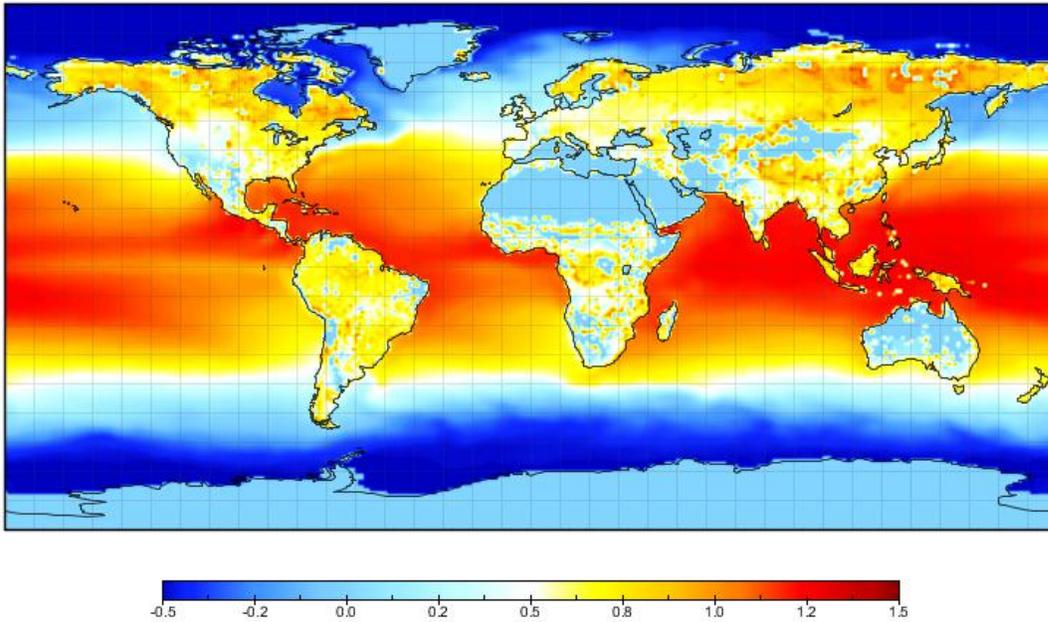
1241



1242

1243 Figure 5. Global $\delta^{13}\text{C}$ distribution over land (annual flux-weighted average in 2003).

1244



1245

1246

1247

Figure 6. Disequilibria between ^{13}C fluxes to and from the land or ocean surface in 2000. At the land

1248

surface, the disequilibrium is the difference between photosynthetic and respiratory discriminations

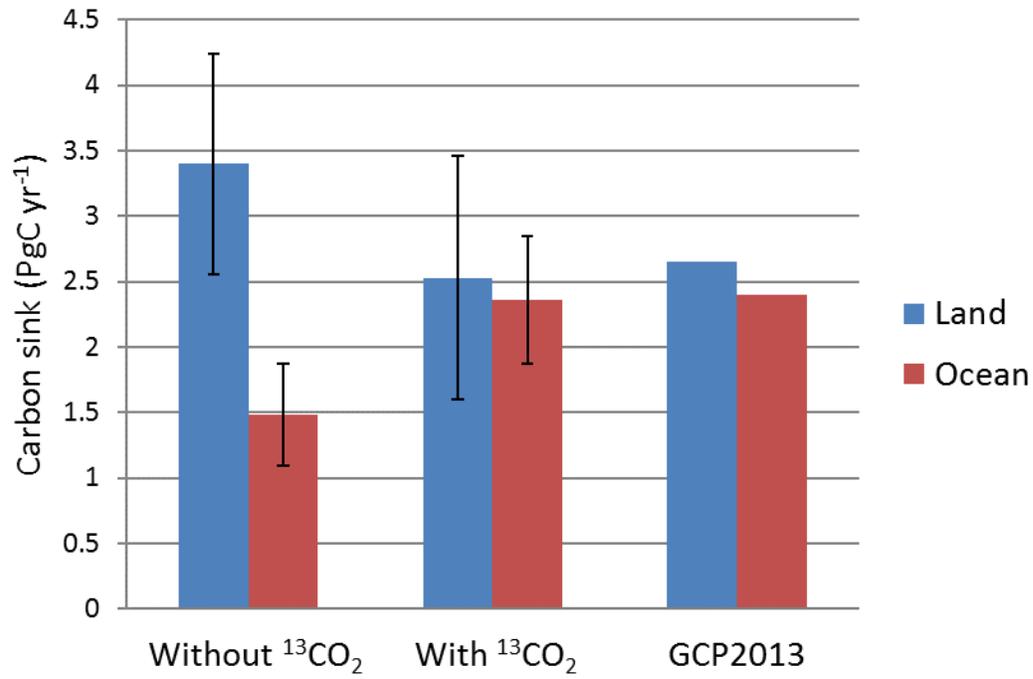
1249

against ^{13}C , and at the ocean surface, it is the difference in ^{13}C discrimination between the one-way

1250

diffusive downward and upward fluxes.

1251

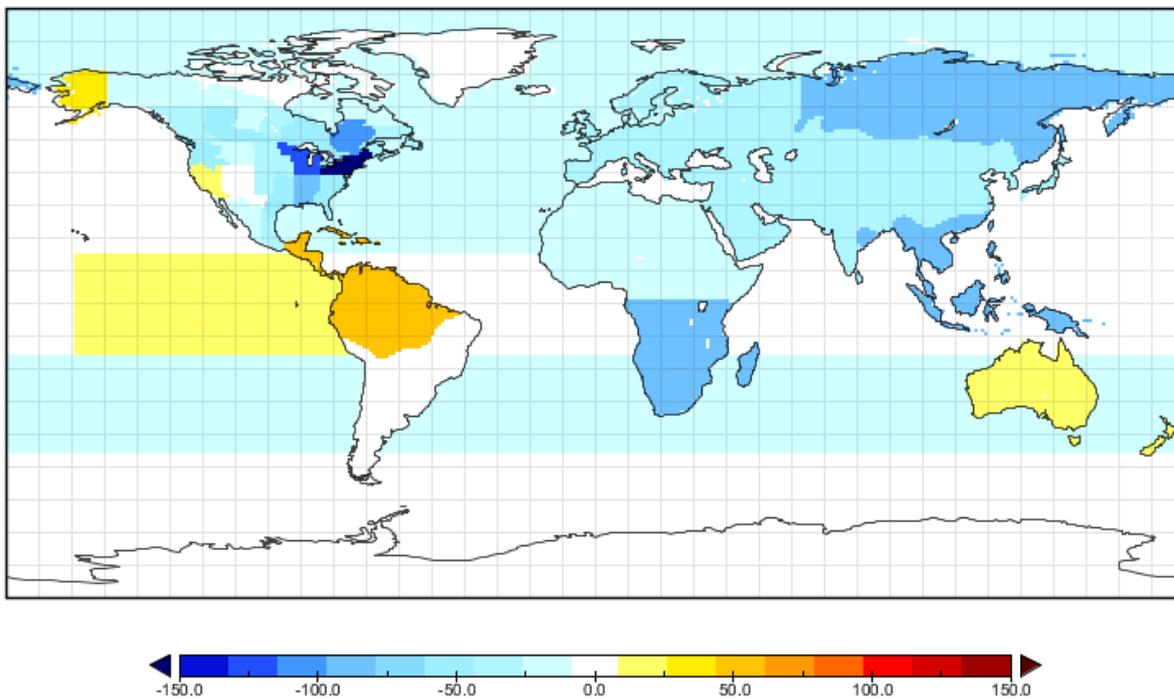


1252

1253 Figure 7. Comparison of land and ocean carbon sinks derived from inversions with and without the
 1254 ¹³CO₂ constraint against the Global Carbon Project results (*Le Quéré et al., 2013*).

1255

1256



1257

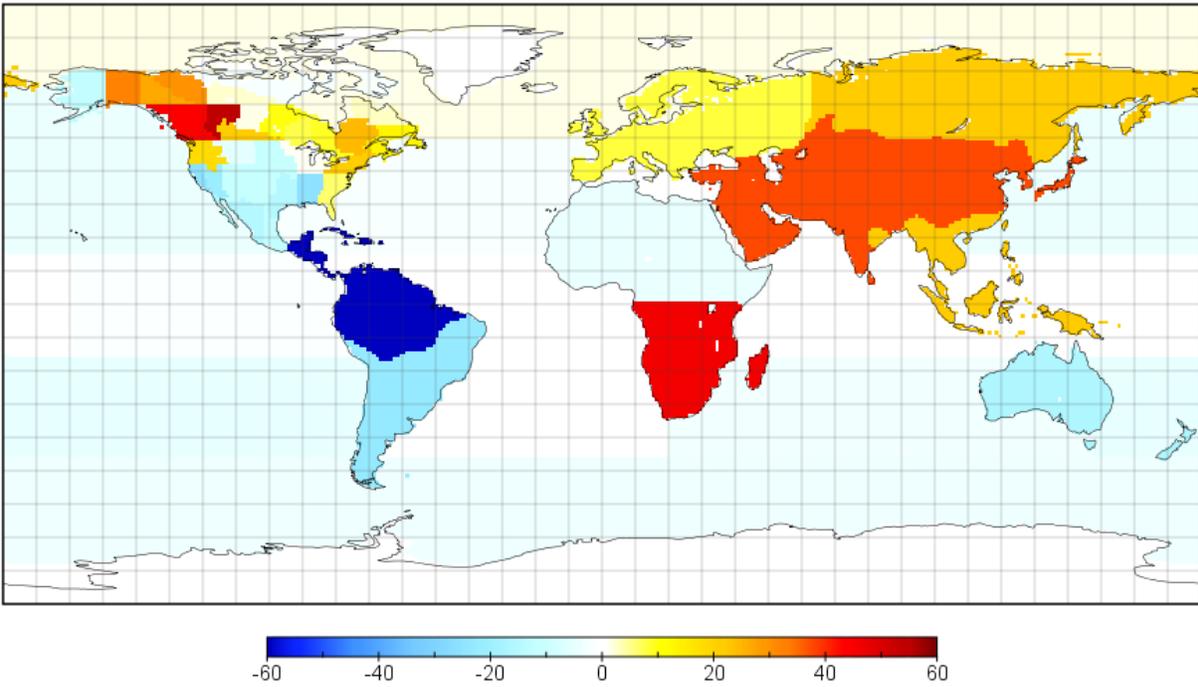
1258 Figure 8. Global distribution of inverted CO₂ flux using CO₂ data only ((gC m⁻²y⁻¹, 2002-2004
1259 average).

1260

1261

1262

1263



1264

1265 Figure 9. Difference of the inverted CO₂ flux between using CO₂ + ¹³CO₂ data and using CO₂ data
1266 only (gC m⁻² y⁻¹, 2002-2004 average).

1267

1268

1269

1270

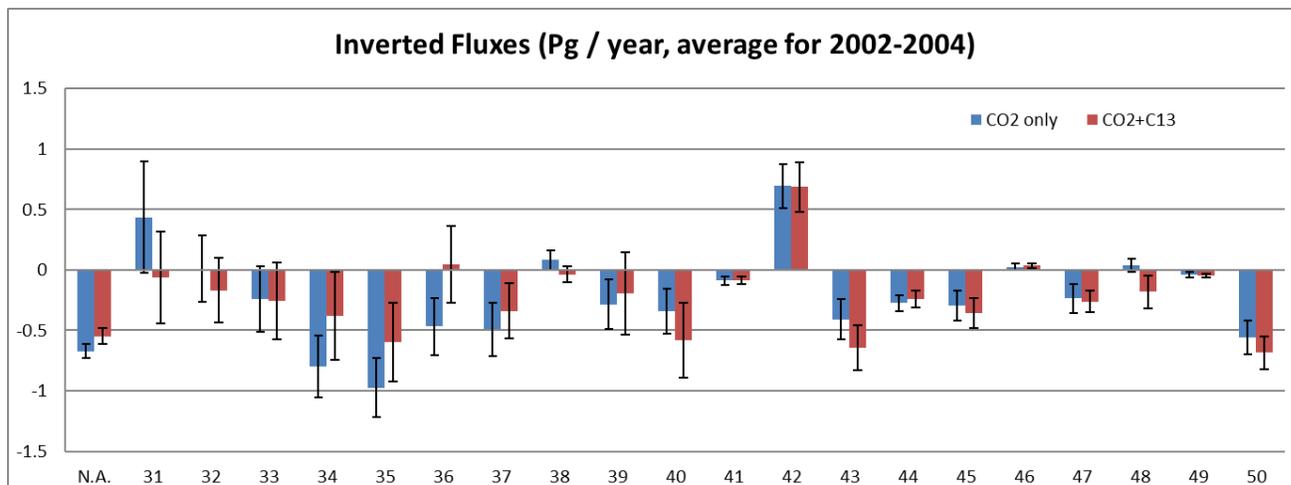
1271

1272

1273

1274

1275

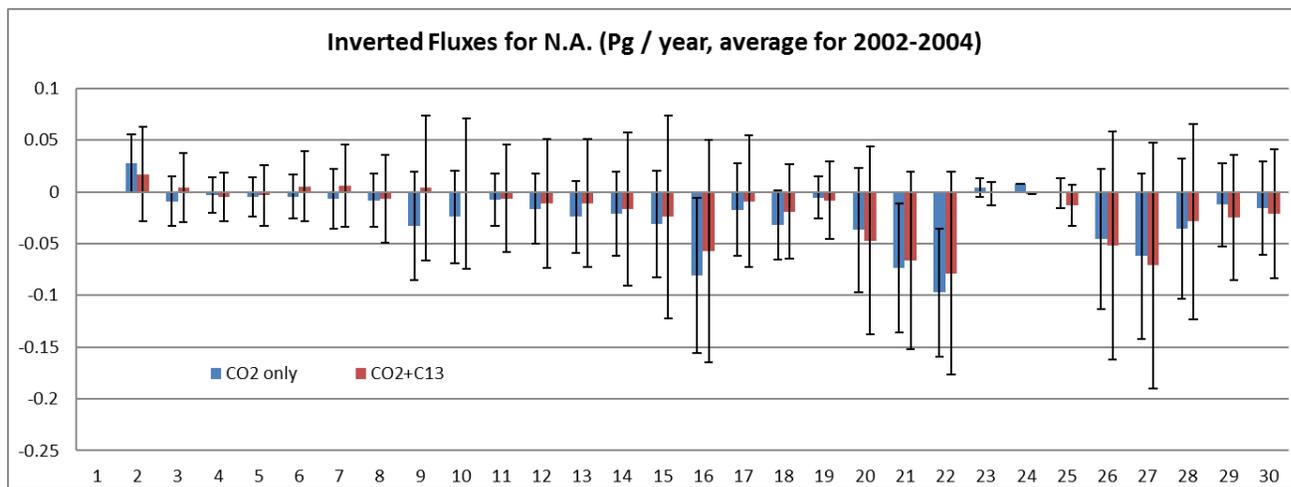


1276

1277 Figure 10. Comparison between inversion results with and without $^{13}\text{CO}_2$ constraint for 21 regions of
1278 the globe for the periods of 2002-2004.

1279

1280
1281

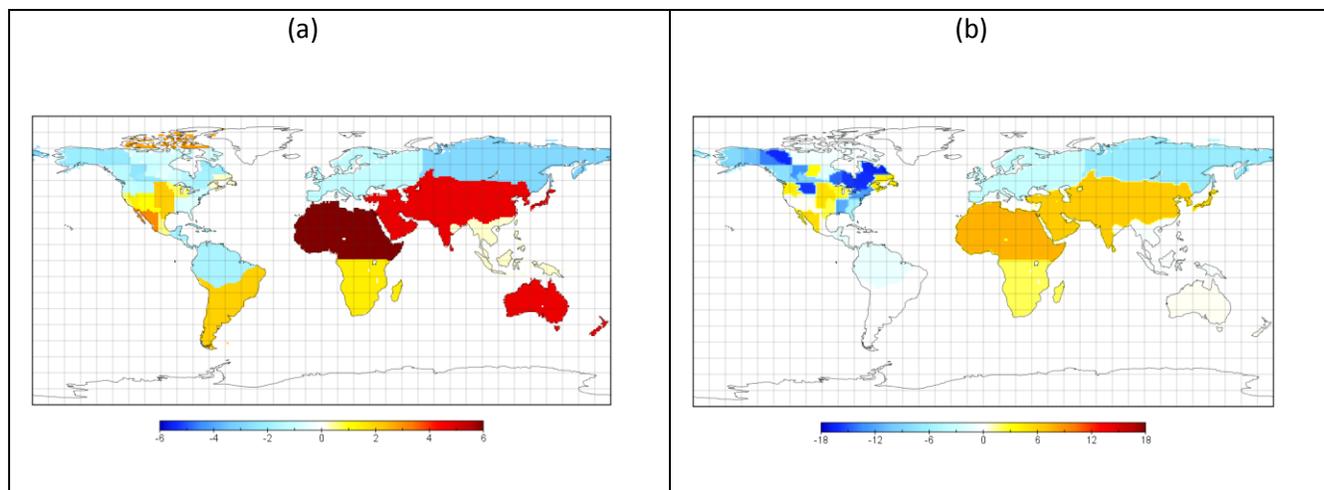


1282

1283 Figure 11. Comparison between inversion results with and without $^{13}\text{CO}_2$ constraint for 30 regions in
1284 North America.

1285

1286

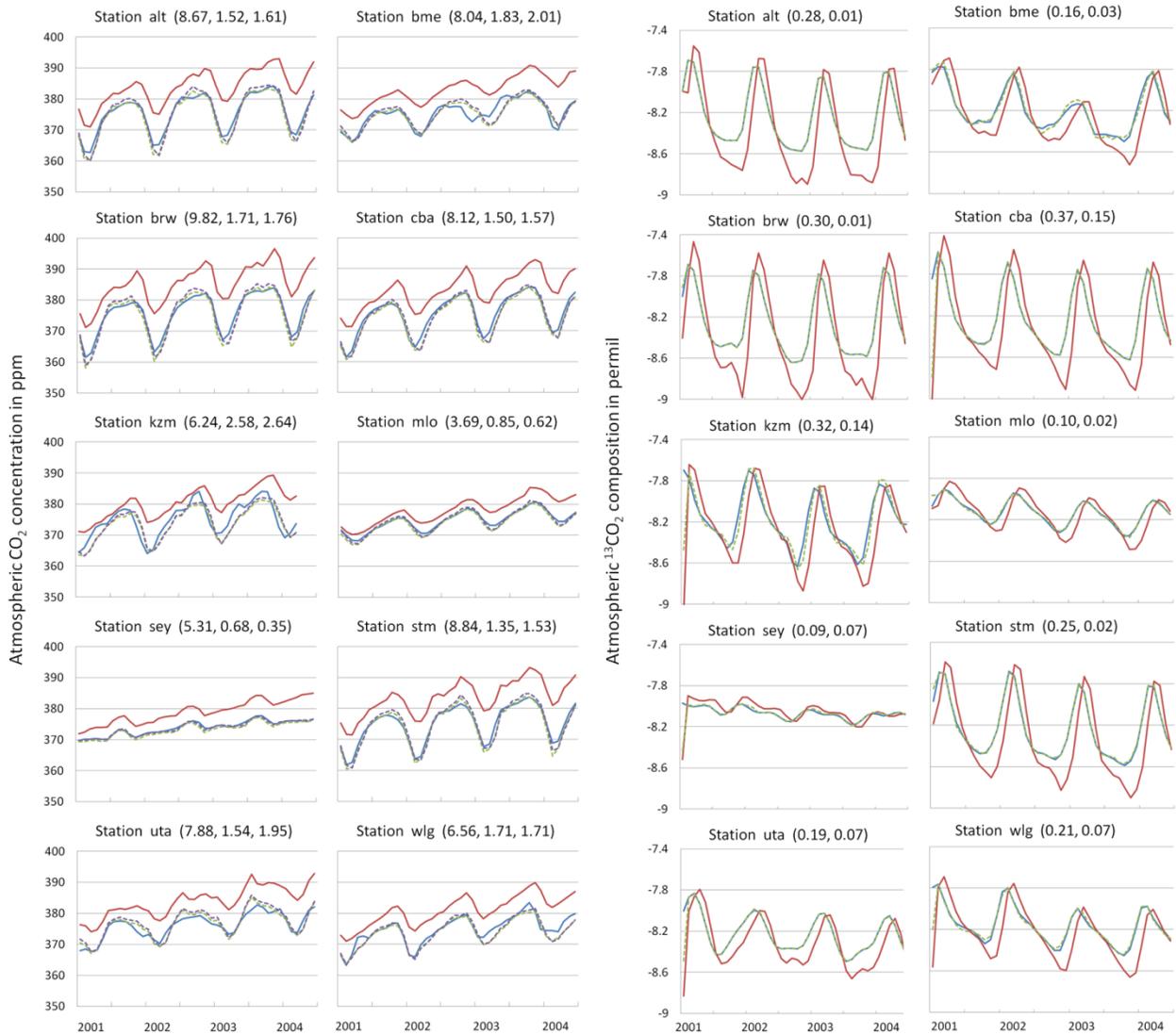


1287

1288 Figure 12. (a) Difference in ϵ_{iph} (‰) and (b) the inverted CO_2 flux ($\text{gC m}^{-2}\text{y}^{-1}$) between Case I and
 1289 Case II, i.e. Case I minus Case II. See Section 2.1.3 for the description of these cases.

1290

1291

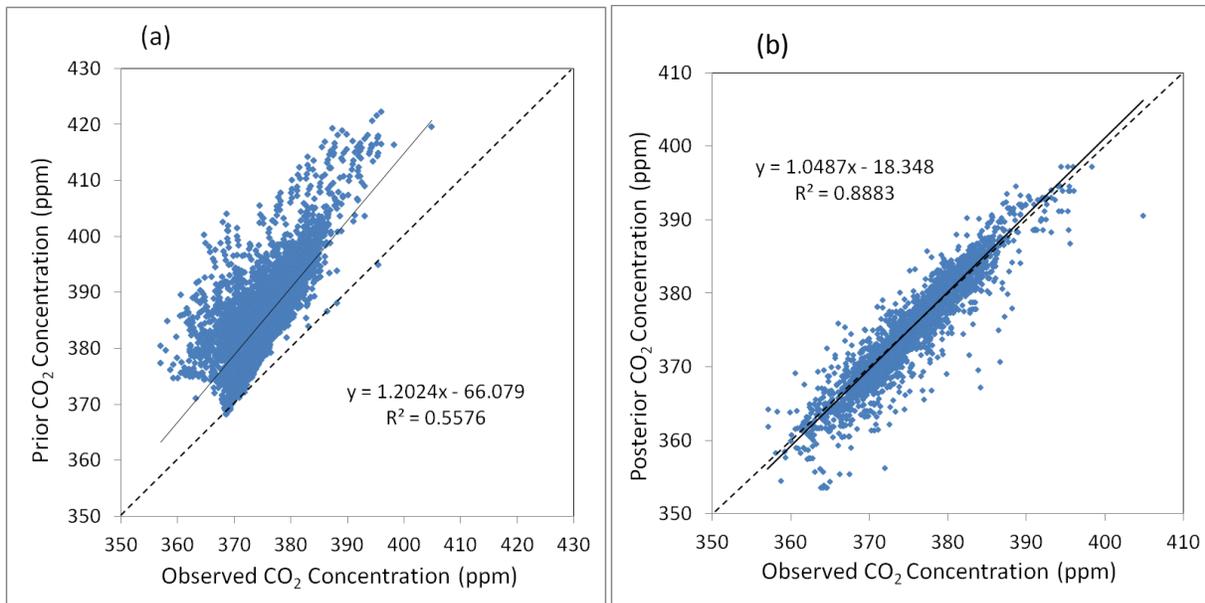


1293

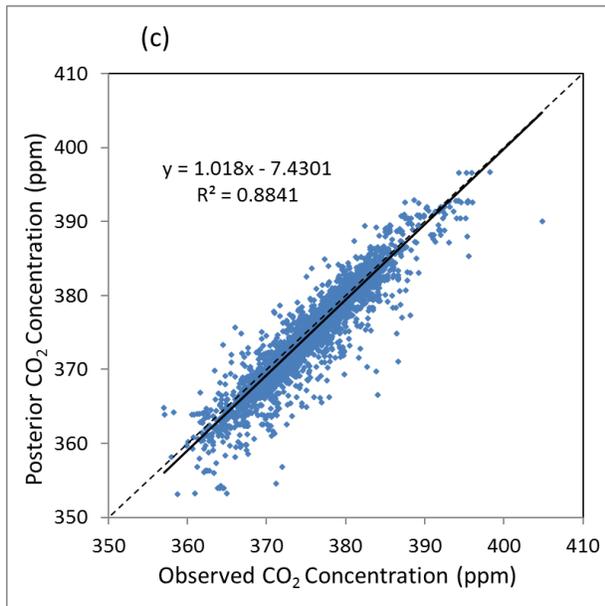
1294 Figure 13. Left panel: comparison of CO₂ concentrations calculated using the prior flux (solid red)
 1295 and from CO₂-only inversion (dashed purple) and joint inversion (dashed green) against observations
 1296 (blue) at 10 randomly selected stations from different regions. The header of each plot indicates the
 1297 station ID and the root mean square difference (RMSD) for the prior, joint and CO₂-only inversions
 1298 against observations. Right panel: comparison of ¹³CO₂ composition from the prior (solid red) and
 1299 joint inversion (dashed green) against observations (blue). The header of each plot indicates the
 1300 station ID and RMSD of the prior and the joint inversion against observations.

1301

1302



1303



1304

1305

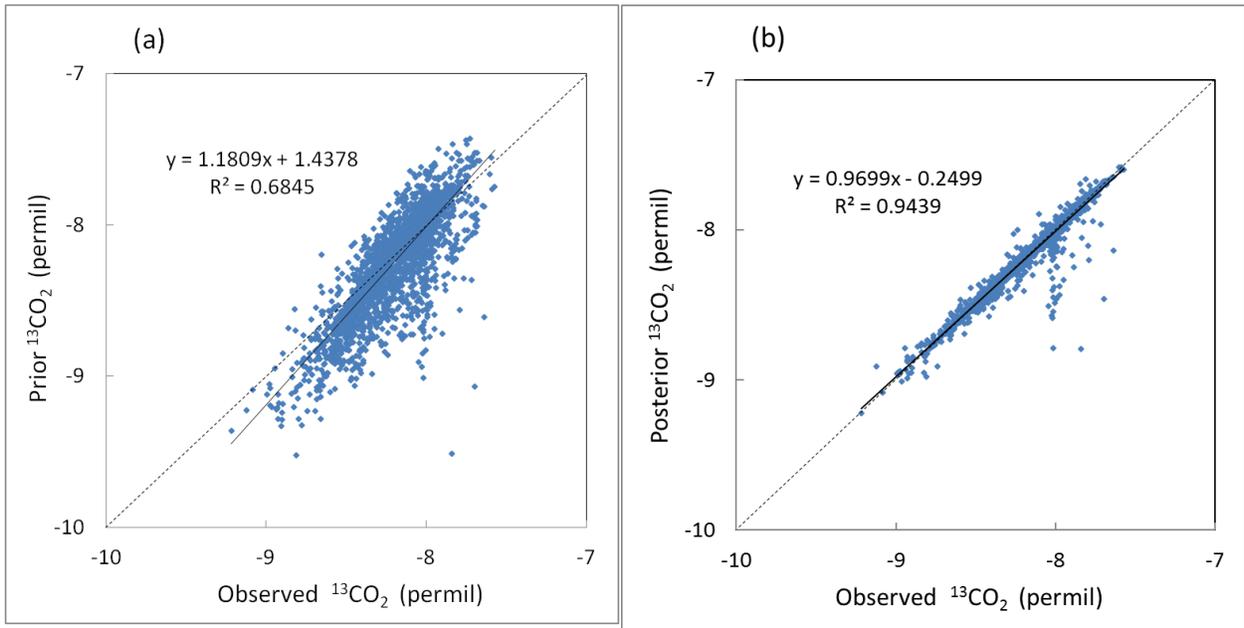
1306 Figure 14. CO₂ concentrations from (a) prior, (b) posterior from the CO₂-only inversion, and (c)
 1307 posterior from the joint inversion in comparison with observations. The prior concentration is
 1308 obtained through transport modeling with prior CO₂ fluxes from the terrestrial ecosystems, oceans,
 1309 fossil fuel emission, and biomass burning.

1310

1311

1312

1313



1314

1315

1316 Figure 15. Comparison of prior (a) and posterior (b) $^{13}\text{CO}_2$ compositions with observations. The prior
1317 composition is obtained through transport modeling with prior $^{13}\text{CO}_2$ fluxes from the terrestrial
1318 ecosystems, oceans, fossil fuel emission, and biomass burning, and the posterior composition is
1319 obtained with the CO_2 - $^{13}\text{CO}_2$ joint inversion (Case I).

1320

1321

1322

1323

1324

1325