1 **Global Soil Consumption of Atmospheric Carbon Monoxide:**

2 **An Analysis Using a Process-Based Biogeochemistry Model**

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Abstract: Carbon monoxide (CO) plays an important role in controlling the oxidizing capacity of the atmosphere by reacting with OH radicals that affect atmospheric methane (CH_4) dynamics. We develop a process-based biogeochemistry model to quantify CO exchange between the soil and the atmosphere at the global scale. The model is parameterized using CO flux data from the field and laboratory experiments for eleven representative ecosystem types. The model is then extrapolated to the global terrestrial ecosystems. Global soil gross consumption, gross production, and net flux of the atmospheric CO are estimated to be 132-154, 29-36 and 102-119 Tg CO yr^{-1} (1Tg = 10^{12} g), respectively, assuming a constant spatially distributed atmospheric CO concentration (~128 ppbv) during the 20th century. When satellite-based atmospheric CO concentration data are used, our estimates of the soil gross consumption are 180-197 Tg CO yr^{-1} in the period of 2000-2013. Tropical evergreen forest, savanna and deciduous forest areas are the largest sinks at 93 Tg CO yr^{-1}. Soil CO gross consumption is sensitive to air temperature and atmospheric CO concentration while gross production is sensitive to soil organic carbon (SOC) stock and air temperature. Under future climate scenarios, the soil gross consumption, gross production and net flux of CO will increase at 0.15-1.23, 0.04-0.3 and 0.12-0.94 Tg CO yr^{-2} during 2014-2100, reaching 162-194, 36-44, and 126-150 Tg CO yr^{-1} by the end of the 21st century, respectively. Areas near the equator, Eastern US, Europe and eastern Asia will be the largest sinks due to optimum soil moisture and high temperature. The annual global soil net flux of atmospheric CO is primarily controlled by air temperature, soil temperature, SOC and atmospheric CO concentrations, while its monthly variation mainly determined by air temperature, precipitation, soil temperature and soil moisture. Our process-based soil CO dynamics model and analysis shall benefit the modeling of the global climate and atmospheric chemistry.

1. Introduction

Carbon monoxide (CO) plays an important role in controlling the oxidizing capacity of the atmosphere by reacting with OH radicals (Logan et al., 1981; Crutzen, 1987; Khalil & Rasmussen, 1990; Prather et al., 1995; Prather & Ehhalt, 2001). CO in the atmosphere can directly and indirectly influence the fate of critical greenhouse

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39 gases such as methane (CH_4) and ozone (O_3) (Logan et al., 1981; Crutzen & Gidel,

40 1983; Guthrie, 1989; Khalil & Rasmussen, 1990; Lu & Khalil, 1993; Daniel & Solomon,

41 1998; Prather & Ehhalt, 2001; Tan and Zhuang, 2012). Although CO itself absorbs only

42 a limited amount of infrared radiation from the Earth, the cumulative indirect radiative

43 forcing of CO may be even larger than that of the third powerful greenhouse gas, nitrous

44 oxide (N_2O, Myhre et al., 2013). Current estimates of global CO emissions from both

45 anthropogenic and natural sources range from 1550 to 2900 Tg CO yr^{-1}, which are

46 mainly from anthropogenic and natural direct emissions and from the oxidation of

47 methane and other Volatile Organic Compounds (VOC) (Prather et al., 1995; Khalil et

48 al., 1999; Bergamaschi et al., 2000; Prather & Ehhalt, 2001, Stein et al., 2014).

49 Chemical consumption of CO by atmospheric OH and the biological consumption of CO

50 by soil microbes are two major sinks of the atmospheric CO (Conrad, 1988; Lu & Khalil,

51 1993; Prather et al., 1995; Prather & Ehhalt, 2001; Yonemura et al., 2000; Whalen &

52 Reeburgh, 2001).

53   Soils are globally considered as a major sink for CO due to microbial activities

54 (Conrad and Seiler, 1982; Potter et al., 1996; Whalen and Reeburgh, 2001; King and

55 Weber, 2007). A diverse group of soil microbes including carboxydotrophs,

56 methanotrophs and nitrifiers are capable of oxidizing CO (Ferenci et al., 1975; Jones

57 and Morita, 1983; Bender and Conrad, 1994; King and Weber, 2007). Annually, 10-25%

58 of CO emissions were consumed by soils (Sanhueza et al., 1998; Khalil et al., 1999;

59 King, 1999a; Bergamaschi et al., 2000; Prather & Ehhalt, 2001; Chan & Steudler, 2006).

60 Potter et al. (1996) reported the global soil consumption be 16-50 Tg CO yr^{-1}, by using a

61 single box model approach over the upper 5 cm of soils. Other estimates showed large

62 ranges using simple assumptions, such as 115–230 Tg CO yr^{-1} based on a constant

63 dry deposition velocity (the uptake rate divided by the CO concentration) of 0.03 cm s^{-1}

64 (Sanhueza et al., 1998); 300 Tg CO yr^{-1} using the same constant deposition velocity

65 and zero deposition velocity value in deserts and areas with monthly mean

66 temperatures below 0 °C with different approaches (Bergamaschi et al., 2000); 190-580

67 Tg CO yr^{-1} using empirical approaches with a higher probability for lower values (King,

68 1999). Besides, reported CO dry deposition velocities (0 to 0.004m s^{-1}) for vegetated

69 surfaces based on measurements are relatively low compared with other substances

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70 (King, 1999a; Castellanos et al., 2011). To date, there are still large uncertainties in

71 estimating soil CO consumption, ranging from 15 to 640 Tg CO yr^{-1}. Although soil CO

72 consumption and its environmental controls have been heavily studied, the impacts of

73 long-term changes in climate and human activities on the atmosphere-biosphere CO

74 exchange are still not clear (King & Weber, 2007; Vreman et al., 2011; He and He, 2014;

75 Pihlatie et al., 2016). Moreover, production of CO has been widely found in soils, plant

76 roots, living and degrading plant materials and degrading organic matter (Pihlatie et al.,

77 2016). CO production is dominantly due to abiotic processes such as thermal- and

78 photo-degradation of organic matter or plant material (Conrad and Seiler, 1985b; Tarr et

79 al., 1995; Schade et al., 1999; Derendorp et al., 2011; Lee et al., 2012; van Asperen et

80 al., 2015; Fraser et al., 2015, Pihlatie et al., 2016), except for a few cases of anaerobic

81 formations. Photo-degradation includes direct photo-degradation due to absorbing

82 radiation by light-absorbing molecules and indirect photo-degradation due to radiation

83 energy transferring to non-light-absorbing molecules (King et al., 2012). Thermal-

84 degradation is identified as the temperature-dependent degradation of carbon in the

85 absence of radiation and possibly oxygen (Derendorp et al., 2011; Lee et al., 2012; van

86 Asperen et al., 2015; Pihlatie et al., 2016). Previous field and laboratory studies on the

87 role of direct or indirect abiotic degradation showed very contrasting results, primarily

88 due to the challenge of separation between CO formation through thermal-degradation

89 and photo-degradation, because they can both occur simultaneously and the indirect

90 photo-degradation may occur even without solar radiation if thermal energy is suitable

91 (Lee et al., 2012).

92 Little focus has been placed so far on the role of net CO budget (including soil

93 CO consumption and production) in global climate modeling. Most top-down models

94 apply a dry deposition scheme based on the resistance model of Wesely (1989). Such

95 schemes give a wide range of dry deposition velocities (Stevenson et al., 2006). Only a

96 few models (MOZART-4, Emmons et al., 2010; CAM-chem, Lamarque et al., 2012)

97 have extended their dry deposition schemes with a parameterization for CO and H$_2$

98 uptake by oxidation from soil bacteria and microbes following the work of Sanderson et

99 al. (2003), which itself was based on extensive measurements from Yonemura et al.

100 (2000). Potter et al. (1996) developed a bottom-up model to simulate CO consumption

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101   and production at the global scale. This model is a single box model, only considers top

102   5cm depth of soil and does not have explicit microbial factors, which might have

103   underestimated CO consumption (Potter et al., 1996; King, 1999a). Current bottom-up

104   CO modeling approaches are mostly based on a limited number of CO *in situ*

105   observations or laboratory studies to quantify regional and global soil consumption

106   (Potter et al., 1996; Sanhueza et al., 1998; Khalil et al., 1999; King, 1999a;

107   Bergamaschi et al., 2000; Prather & Ehhalt, 2001). To our knowledge, no detailed

108   process-based model of soil-atmospheric exchange of CO has been published in the

109   recent 15 years. One reason is an incomplete understanding of biological processes of

110   CO emission and uptake (King & Weber, 2007; Vreman et al., 2011; He and He, 2014;

111   Pihlatie et al., 2016). Another reason is a lack of long-term CO flux measurements for

112   different ecosystem types to calibrate and evaluate the models. CO flux measurements

113   are mostly from short-term field observations or laboratory experiments (e.g. Conrad

114   and Seiler, 1985a; Funk et al., 1994; Tarr et al., 1995; Zepp et al., 1997; Kuhlbusch et

115   al., 1998; Moxley and Smith, 1998; Schade et al., 1999; King and Crosby, 2002; Varella

116   et al., 2004; Lee et al., 2012; Bruhn et al., 2013; van Asperen et al., 2015). The first

117   study to report long-term and continuous field measurements of CO flux over grassland

118   using a micrometeorological eddy covariance (EC) method are in Pihlatie et al. (2016).

119   Aiming to improve the understanding of processes associated with land-

120   atmosphere CO exchange and to quantify global soil CO budget for the 20th and 21st

121   centuries, we developed a CO dynamics module (CODM) embedded in a process-

122   based biogeochemistry model, the Terrestrial Ecosystem Model (TEM) (Zhuang et al.,

123   2003, 2004, 2007). CODM was then calibrated and evaluated using laboratory

124   experiments and field measurements for different ecosystem types. We then used the

125   atmospheric CO concentration data from MOPITT (Gille, 2013) to drive our model from

126   2000 to 2013.   We conducted century-long simulations of 1901-2100, using the

127   atmospheric CO concentrations estimated withan empirical function (Badr & Probert,

128   1994; Potter et al., 1996). We also evaluated the effects of multiple forcings on global

129   CO consumption and production estimates, including the changes of climate and

130   atmospheric CO concentrations at the global scale.

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2. Method

2.1 Overview

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134   We first developed a daily soil CO dynamics module (CODM) that considers: (1)
135 soil-atmosphere CO exchange and diffusion process between soil layers, (2)
136 consumption by soil microbial oxidation, (3) production by soil chemical oxidation, and
137 (4) the effects of temperature, soil moisture, soil CO substrate and surface atmospheric
138 CO concentration on these processes. Second, we used observed soil temperature and
139 moisture to evaluate TEM hydrology module and soil thermal module in order to
140 estimate soil physical variables correctly. Then we used results of laboratory
141 experiments and CO flux measurements to parameterize the model and calibrate the
142 model using the Shuffled Complex Evolution (SCE-UA) method (Duan et al., 1993).
143 Finally, the model was extrapolated to the global scale at a 0.5° by 0.5° resolution. We
144 conducted three sets of model experiments to investigate the impact of climate and
145 atmospheric CO concentrations on soil CO dynamics: 1) 1901-2013 with constant
146 atmospheric CO concentrations estimated from an empirical function; 2) 2000-2013 with
147 MOPITT satellite atmospheric CO concentration data; and 3) 2014-2100 with the same
148 constant atmospheric CO concentrations as 1) and three future climate scenarios.

149

2.2 Carbon Monoxide Dynamics Module (CODM)

151   Embedded in TEM (Figure 1), CODM is mainly driven by: (1) soil organic carbon
152 availability based on a carbon and nitrogen dynamics module (CNDM) (Zhuang et al.,
153 2003); (2) soil temperature profile from a soil thermal module (STM) (Zhuang et al.,
154 2001, 2003); and (3) soil moisture profile from a hydrological module (HM) (Bonan,
155 1996; Zhuang et al, 2004). Net exchange of CO between the atmosphere and soil is
156 determined by the mass balance. According to previous studies, we separate active
157 soils (top 30cm) for CO consumption and production into 1 cm thick layers (King, 1999a,
158 1999b; Whalen & Reeburgh, 2001; Chan & Steudler, 2006). Between the soil layers, the
159 changes of CO concentrations are calculated by:

$$\frac{\partial(C(t,i))}{\partial t} = \frac{\partial}{\partial z}\left(D(t,i)\frac{\partial\big(C(t,i)\big)}{\partial z}\right) + P(t,i) - O(t,i) \qquad (1)$$

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160 Where $C(t,i)$ is the CO concentration in layer i and time t, units are mg m^{-3}. z is the

161 thickness of layer i. $D(t,i)$ is the diffusion coefficient for layer i, units are m^2 s^{-1}. $P(t,i)$

162 is the CO production rate and $O(t,i)$ is CO consumption rate due to oxidation. The units

163 of $P(t,i)$ and $O(t,i)$ are mg m^{-3} s^{-1}. $D(t,i)$ is calculated using the method from Potter et

164 al. (1996), equation (2) to (4), which is the function of soil temperature, soil texture and

165 soil moisture. The upper boundary condition is specified as the atmospheric CO

166 concentration, which is estimated by an empirical function of latitude (Potter et al., 1996)

167 or directly measured by the MOPITT satellite during 2000-2013. The lower boundary

168 condition is assumed to have no diffusion exchange with the layer underneath. This

169 partial differential equation (PDE) is solved using the Crank-Nicolson method for less

170 time-step-sensitive solution.

171   CO consumption is modeled as an aerobic process occurring in unsaturated soil

172 pores, which is estimated as:

$$O(t,i) = V_{max} \cdot f_1\big(C(t,i)\big) \cdot f_2\big(T(t,i)\big) \cdot f_3\big(M(t,i)\big) \quad (2)$$

173 Where V_{max} is the specific maximum oxidation rate, ranging from 0.3 to 11.1 µg CO g^{-1}

174 h^{-1} (Whalen & Reeburgh, 2001). f_i are functions calculating CO concentration $C(t,i)$,

175 temperature $T(t,i)$ and moisture $M(t,i)$ influences on CO soil consumption.

176 Considering CO consumption as the result of microbial activities, we calculate

177 $f_1\big(C(t,i)\big)$, $f_2\big(T(t,i)\big)$ and $f_3\big(M(t,i)\big)$ in a similar way as Zhuang et al. (2004):

$$f_1\big(C(t,i)\big) = \frac{C(t,i)}{C(t,i) + k_{CO}} \quad (2.1)$$

$$f_2\big(T(t,i)\big) = Q_{10}^{\frac{T(t,i)-T_{ref}}{10}} \quad (2.2)$$

$$f_3\big(M(t,i)\big) = \frac{(M(t,i) - M_{min})(M(t,i) - M_{max})}{(M(t,i) - M_{min})(M(t,i) - M_{max}) - (M(t,i) - M_{opt})^2} \quad (2.3)$$

178 Where $f_1\big(C(t,i)\big)$ is a multiplier that enhances oxidation rate with increasing soil CO

179 concentrations using a Michaelis-Menten function with a half-saturation constant k_{CO},

180 ranging from 5 to 51 µl CO l^{-1} (Whalen & Reeburgh, 2001); $f_2\big(T(t,i)\big)$ is a multiplier that

181 enhances CO oxidation rates with increasing soil temperature using a Q10 function with

182 Q_{10} coefficients (Whalen & Reeburgh, 2001). T_{ref} is the reference temperature, units

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183   are °C (Zhuang et al., 2004, 2013). $f_3(M(t,i))$ is a multiplier to estimate the biological

184   limiting effect that diminishes CO oxidation rates if the soil moisture is not at an optimum

185   level (M_{opt}). M_{min}, M_{max} and M_{opt} are the minimum, maximum and optimum volumetric

186   soil moistures of oxidation reaction, respectively. Equation (2.2) will overestimate CO

187   consumption at higher temperature because CO consumption has an optimum

188   temperature and it will decrease at higher temperatures. However, the CO consumption

189   is constrained by CO production, and equation (1) is used to represent this constraint.

190   We model the CO production rate ($P(t,i)$) as a process of chemical oxidation

191   constrained by soil organic carbon (SOC) decay (Conrad and Seiler,1985; Potter et al.

192   1996; Jobbagy & Jackson, 2000; van Asperen et al., 2015):

$$P(t,i) = P_r(t,i) \cdot E_{SOC} \cdot C_{SOC}(t) \cdot F_{SOC} \qquad (3)$$

193   Where $P_r(t,i)$ is a reference soil CO production rate which has been normalized to rate

194   at reference temperature, which is affected by soil moisture and soil temperature

195   (Conrad and Seiler,1985; van Asperen et al., 2015). E_{SOC} is an estimated nominal CO

196   production factor of 3.5 ± 0.9 X 10^{-9} mg CO m^{-2} s^{-1} per g SOC m^{-2} (to 30 cm surface soil

197   depth) (Potter et al., 1996). $C_{SOC}(t)$ is a SOC content in mg m^{-2}, which is provided by

198   CNDM module in TEM. F_{SOC} is a constant fraction of top 20cm SOC compared to total

199   amount of SOC, which is 0.33 for shrubland areas, 0.42 for grassland areas and 0.50

200   for forest areas, respectively (Jobbagy & Jackson, 2000). $P_r(t,i)$ is calculated as:

$$P_r(t,i) = \exp\left(f_4\big(M(t,i)\big) \cdot Ea_{ref}/R \cdot \left(\frac{1}{273.15 + PT_{ref}} - \frac{1}{T(t,i) + 273.15} \right) \right) \qquad (3.1)$$

$$f_4\big(M(t,i)\big) = \frac{PM_{ref}}{M(t,i) + PM_{ref}} \qquad (3.2)$$

201   Where equation (3.1) is derived from Arrhenius equation for chemical reactions and

202   normalized using the reference temperature PT_{ref}. Ea_{ref}/R is the reference activation

203   energy divided by gas constant R, units are K. $f_4\big(M(t,i)\big)$ is the multiplier that reduces

204   activation energy using an regression approach based on laboratory experiment of

205   moisture influences on CO production (Conrad and Seiler,1985). PM_{ref} is the reference

206   volumetric soil moisture, ranging from 0.01 to 0.5 volume/volume (v/v). We assume

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207 thermal-degradation as the main CO producing process since lack of photo-degradation

208 data and hard to distinguish photo-degradation from observations. In order to reduce

209 the bias from thermal-degradation to total abiotic degradation, the equation (3.1) is

210 parameterized by comparing with total production rate. For instance, $P_r(t, i)$ calculation

211 can perfectly fit the experiment results in Van Asperen et al., 2015 with proper

212 PT_{ref}(18°C), Ea_{ref}/R(14000 K) and PM_{ref}(0.5 v/v).

213

214 **2.3 Model Parameterization and Extrapolation**

215   The model parameterization was conducted in two steps: 1) Thermal and

216 hydrology modules embedded in TEM were revised, calibrated and evaluated by

217 running model with corresponding local meteorological or climatic data at 4

218 representative sites, including boreal forest, temperate forest, tropical forest and

219 savanna (Table 1, site No.1 to 4, Figure 2) to minimize model data misfit in terms of soil

220 temperature and moisture. 2) CODM module was parameterized by running TEM for

221 observational periods with the corresponding local meteorological or climatic data at

222 each reference site (Table 1, Figure 3), and using Shuffled Complex Evolution

223 Approach in R language (SCE-UA-R) (Duan et al., 1993) to minimize the difference

224 between simulated and observed net CO flux. Eleven parameters including k_{CO}, V_{max},

225 T_{ref}, Q_{10}, M_{min}, M_{max}, M_{opt}, E_{SOC}, Ea_{ref}/R, PM_{ref} and PT_{ref} are optimized (Table 2). To

226 be noticed, F_{SOC} was not involved in the calibration process. Parameter priors were

227 decided based on previous studies (Conrad & Seiler, 1985; King, 1999b; Whalen &

228 Reeburgh, 2001; Zhuang et al., 2004). SCE-UA-R was used for site No. 6, 8, 10, 11

229 (Table 1). Each site had been run 50 times using SCE-UA-R with 10000 maximum

230 loops for parameter ensemble, and all of them reached stable state before the end of

231 the loops. For wetlands, the only available data is from site No.12. We used trial-and-

232 error method instead to make our simulated results in the range of observed flux rates,

233 with a 10% tolerance. For tropical sites, since tropical savanna vegetation type is a

234 combination type of tropical forest and grassland in our model, we first used Site No. 13

235 to set priors to fit the experiment results with a 10% tolerance and then evaluated by

236 running our model comparing with site No.7 results. Site No. 9 and 5 were used to

237 evaluate our model results for temperate forest and grassland. Besides the observed

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238   climatic and soil property data, we used ERA-Interim reanalysis data from The

239   European Centre for Medium-Range Weather Forecasts (ECMWF) (Dee et al., 2011),

240   AmeriFlux observed meteorology data (http://ameriflux.lbl.gov/) and reanalysis climatic

241   data from Climatic Research Unit (CRU, Harris et al., 2013) to fill the missing

242   environmental data. To sum up, parameters for various ecosystem types in table 2 were

243   the final results of our parameterization. Model parameterization was conducted for

244   ecosystem types including boreal forest, temperate coniferous forest, temperate

245   deciduous forest, and grassland using SCE-UA-R. Tropical forest and wet tundra used

246   a trial-and-error method to adjust parameters letting simulation results best fit the lab

247   data. Due to limited data availability, we assumed temperate evergreen broadleaf forest

248   having the same parameters as temperate deciduous forest.

249

250   **2.4 Data Organization**

251   To get spatially and temporally explicit estimates of CO consumption, production

252   and net flux at the global scale, we used the data of land cover, soils, climate and leaf

253   area index (LAI) from various sources at a spatial resolution of 0.5° latitude X 0.5°

254   longitude to drive TEM. The land cover data include potential vegetation distribution

255   (Melillo et al., 1993) and soil texture (Zhuang et al., 2003), which were used to assign

256   vegetation- and texture-specific parameters to each grid cell.

257   For the simulation of the period 1901-2013, monthly air temperature, precipitation,

258   clouds fraction and vapor pressure data sets from CRU were used to estimate the soil

259   temperature, soil moisture and SOC with TEM (Figure 4). Monthly LAI data from TEM

260   were required to simulate soil moisture (Zhuang et al., 2004). During this period time,

261   we used an empirical function of latitude, which was derived from the observed

262   latitudinal distribution of tropospheric carbon monoxide (Badr and Probert, 1994) to

263   calculate CO surface concentrations (equation (7), Potter et al., 1996):

$$C_{CO,air} = 82.267856 + 0.8441503L + 1.55934 \times 10^{-2}L^2 + 2.37 \times 10^{-5}L^3 - 2.3 \times 10^{-6}L^4$$

264   Where $C_{CO,air}$ is the derived surface CO concentration (ppbv), *L* represents

265   latitude which is negative degrees for southern hemisphere and positive degrees for

266   northern hemisphere. We also used the transient atmospheric CO data from MOPITT

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267 satellite during 2000-2013 (Figure 5). We averaged day-time and night-time monthly

268 mean retrieved CO surface level 3 data (variables mapped on 0.5° latitude X 0.5°

269 longitude grid scales with monthly time step, Gille, 2013) to represent the CO surface

270 concentration level in each month. The missing pixels were fixed by the average of

271 pixels which had values and were inside 1.5 times of the distance between this missing

272 pixel and the nearest pixel with values. These global mean values shown in Figure 5 do

273 not include ocean surfaces, thus there are differences between our surface CO

274 concentration results and Yoon and Pozzer's report in 2014, which is as low as 99.8ppb.

275 From 2014 to 2100, we used Intergovernmental Panel on Climate Change (IPCC) future

276 climate scenarios from Representative Concentration Pathways (RCPs) climate forcing

277 data sets RCP2.6, RCP4.5 and RCP8.5 (Figure 6). Since RCPs did not have water

278 vapor pressure data, we use the specific humidity and sea level air pressure from the

279 RCPs and elevation of surface to estimate the monthly surface vapor pressure data

280 (Seinfeld & Pandis, 2006).

281

282 **2.5 Model Experiment Design**

283   We conducted two core simulations and eight sensitivity test simulations in

284 historical period. The two core simulations were driven with CO surface concentrations

285 estimated from an empirical function of latitude (experiment E1) for the period 1901-

286 2013 and with transient CO surface concentrations from MOPITT satellite data

287 (experiment E2) for the period 2000-2013, respectively. Eight sensitivity simulations

288 were driven with constant CO surface concentrations ± 30%, SOC ±30%, precipitation

289 ±20% and air temperature ± 3°C for each pixel during 1999-2000 (E3). For the 21st

290 century, we conducted simulations driven with climate data of RCP2.6, RCP4.5 and

291 RCP8.5 to examine the responses of CO flux to changing climates (E4).

292

293 **3. Results**

294 **3.1 Site Evaluation**

295   Both the magnitude and variation of the simulated soil temperature and moisture

296 from cold area to warm area compared well to the observations (Figure. 2). The

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297   magnitude of simulated CO flux is highly correlated with the observations (r is about 0.5,

298   p-value < 0.001, Figure 3, a2, b2 ,c2 ,d2). Root mean square error (RMSE) of the

299   simulated CO flux for all sites is below 1.5 mg CO m^{-2} day^{-1}. RMSE for site No. 7 is

300   bigger than 2.0 mg CO m^{-2} day^{-1} when compared with transparent chamber

301   observations. For boreal forest site, we only have 8 acceptable points in 1994 and 1996

302   (Figure 3c2).

303   **3.2 Global Soil CO Dynamics During 1901-2013**

304   For the simulation with constant CO surface concentrations (E1) during 1901-

305   2013, the estimated mean soil CO consumption, production and net flux (positive

306   direction is from soil to atmosphere) are -141, 32 and -108 Tg CO yr^{-1}, respectively. In

307   the long-term simulations, annual soil CO fluxes vary slightly. The annual soil CO

308   consumption, production and net flux vary within 10% during the period (Figure 8a).

309   Consumption is about 4 times larger than production. The highest rates of consumption

310   and production are located in areas close to the equator, and consumption from areas

311   such as eastern US, Europe and eastern Asia also has large rates (>-1000 mg m^{-2} yr^{-1})

312   (Figure 7a, b). Globally soils serve as atmospheric CO sink (Figure 7c). Some areas,

313   such as western US and southern Australia, are CO sources, all of which are grassland

314   or experiencing dry climate. The latitudinal distributions of consumption, production and

315   net flux rates share the same spatial pattern. Around 20°S-20°N and 20-60N° are the

316   largest and second largest areas for production and consumption, while the 45°S-45°N

317   area accounts for nearly 90% of total consumption and production (Figure 9a, Table 3).

318   The Southern and Northern Hemispheres consume 42% and 58% of the total

319   consumption, and produce 41% and 59% of total production, respectively (Table 3).

320   Tropical evergreen forests are the largest sinks, consuming 66 Tg CO yr^{-1}, and tropical

321   savanna and deciduous forest are second and third largest sinks, consuming a total of

322   27 Tg CO yr^{-1} (Table 4). These three ecosystems account for 66% of the total

323   consumption. Tropical evergreen forests are also the largest source of soil CO

324   production, producing 15 Tg CO yr^{-1}, while tropical savanna have a considerable

325   production 6 Tg CO yr^{-1} (Table 4). Moreover, tropical areas, including forested wetlands,

326   forested floodplain and evergreen forests, are most efficient for CO consumption,

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327   ranging from -10 to -12 mg CO m^{-2} day^{-1}. They are also the most efficient for CO

328   production at over 2 mg CO m^{-2} day^{-1} (Table 3).

329   For the simulation with transient atmospheric CO surface concentrations (E2)

330   during 2000-2013, the mean annual global soil consumption increases to -187 Tg CO

331   yr^{-1}, and areas near the equator become large sinks for atmospheric CO together with

332   eastern US, Europe, and eastern Asia (Figure 7) due to the heavy atmospheric CO

333   burden over these areas (Figure 5a). The annual consumption and net flux trends follow

334   the atmospheric CO concentration trends (Figure 5b, Figure 8b), with a small

335   interannual variability (<10%). The latitudinal distributions of soil CO fluxes for E1 and

336   E2 are similar but E2's CO fluxes magnitudes are larger than E1's and around 30°N of

337   E2's distribution shows another peak of CO consumption, due to the high atmospheric

338   CO concentration over eastern Asia (Figure 5a, Figure 9b). The consumption between

339   45°S-45°N increases by 35%, to -137 Tg CO yr^{-1}, which is 73% of the global total

340   annual consumption. Consumption rates of high latitude areas (45°N North) do not

341   change significantly (Figure 7, 9, Table 3), and the annual consumption only increases

342   by 10%, thus the portion of soil CO sinks in northern high latitudes decreases from 12%

343   to 10% of the global total.

344

345   3.3 Global Soil CO Dynamics During 2014-2100

346   Using the constant atmospheric CO, the estimated annual mean soil CO

347   consumptions for the period 2014-2100 are -162, -174 and -194 Tg CO yr^{-1} while

348   estimated annual mean soil productions are 36, 39 and 44 Tg CO yr^{-1} for RCP2.6, 4.5

349   and 8.5 scenarios, respectively. The net fluxes are -118.06, -117.31 and -115.13 Tg

350   CO yr^{-1} at the beginning 10 years of the 21st century, and will reach -127.17, -144.99

351   and -187.25 Tg CO yr^{-1} at the end of the 21st century for RCP2.6, RCP4.5 and RCP8.5

352   scenarios, respectively (Figure 11). Global distribution patterns of CO consumption,

353   production and net flux are similar to the 20th century but there are significant

354   differences among RCP2.6 RCP4.5 and 8.5 scenarios on areas near the equator, flux

355   rates increasing from RCP2.6 to 8.5. Areas near the equator and eastern Asia become

356   big sinks of atmospheric CO, while northeastern US becomes a small source (Figure

357   10). The consumption has relatively fast growth rates during the 21st century (Figure 11).

358 Furthermore, there are significant trends of increasing consumption, production and net

359 flux for nearly all scenarios. The rate ranges of increasing of consumption, production,

360 and net flux are -0.15 to -1.23, 0.04 to 0.3, and -0.12 to 0.94 Tg CO yr^{-2}, respectively

361 (Figure 11). These increasing trends are similar to air temperature increasing trends

362 (Figure 6).

363

364 **4. Discussion**

365 **4.1 Comparison with Other Studies**

366 Previous studies estimated a large range of global CO consumption from -16 to -

367 636 Tg CO yr^{-1}. Our estimates are -132 to -154 Tg CO yr^{-1} for the 20th century and -180

368 to -197 Tg CO yr^{-1} for 2000-2013 using MOPITT satellite CO surface concentration data.

369 Previous studies also provide a large range for CO production from 0 to 7.6 mg m^{-2} day^{-1}

370 ^{1} (reviewed in Pihlatie et al., 2016). Our results showed averaged CO production

371 ranging from 0.01 to 2.29 mg m^{-2} day^{-1}. The large uncertainty of these estimates is

372 mainly due to a different consideration of the microbial activities, the depth of the soil,

373 and the parameters in the model. In contrast to the estimates of -16 to -57 Tg CO yr^{-1}

374 which were based on top 5 cm soils (Potter et al., 1996), our estimates considered

375 30cm soils, just as used in Whalen & Reeburgh (2001). In addition, we used a thinner

376 layer division (1cm each layer) for diffusion process, and used the Crank-Nicolson

377 method solving partial differential equations to avoid time step influences. We also

378 included microbial CO oxidation process to remove the CO from soils and the effects of

379 soil moisture, soil temperature, vegetation type and soil CO substrate on microbial

380 activities. Besides, our soil thermal, soil hydrology and carbon and nitrogen dynamics

381 simulated in TEM provide carbon substrate spatially and temporally for estimating soil

382 CO dynamics (Bonan, 1996; Zhuang et al., 2001, 2003, 2004, 2007). Overall, although

383 a few previous studies have examined the long-term impacts of climate, land use and

384 nitrogen depositions on CO dynamics (Chan & Steudler, 2006, Pihlatie et al., 2016),

385 global prediction of soil CO dynamics still have a large uncertainty.

386

387 **4.2 Major Controls to Soil CO Dynamics**

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388   Eight sensitivity tests have been conducted for the 1999-2000 period, including

389   changing atmospheric CO by ±30%, SOC by ±30%, precipitation by ±30% and air

390   temperature by ±3°C for each pixel (Table 5). Soil CO consumption is most sensitive

391   (changing 29%) to air temperature while production is most sensitive (changing up to

392   36%) to both air temperature and SOC (30%). The net CO fluxes have the similar

393   sensitivities to consumption, because consumption is normally much larger than CO

394   production so that it will determine the dynamics of the net flux. Annual CO consumption,

395   production and net flux follow the change of air temperature (Table 5), which explains

396   the small increasing trends after the 1960s, the significant increasing trend in the 21st

397   century and the large sinks over tropical areas. Besides, a 30% change in precipitation

398   will not lead to large changes in CO flux (< 3%). SOC did not directly influence CO

399   consumption. Increasing SOC led to an increase in soil CO substrate so implying that

400   more CO in soils can be consumed. CO concentrations will only influence the uptake

401   rate and soil CO substrate concentrations, thus influencing the soil CO consumption

402   rate.

403   Annual CO consumption, production and net flux are significantly correlated with

404   air temperature and soil temperature, due to increasing microbial activities (R > 0.91

405   globally). Specifically, annual CO production is strongly correlated with annual mean

406   SOC. The annual mean SOC follows air temperature trends (Figure 4) as CO flux.

407   Consumption has low correlations with annual precipitation and soil moisture, especially

408   at 45°N-45°S (Table 6). The soil moisture is significantly influenced by temperature

409   since increasing temperature would result in higher evapotranspiration. In contrast, the

410   monthly consumption and production are correlated with the precipitation and soil

411   moisture in the Northern Hemisphere (R>0.85), which contains over 53% of the global

412   soil CO consumption (Table 3). Meanwhile, the monthly CO flux is still well correlated

413   with air temperature and soil moisture. Monthly CO flux has low correlations with SOC

414   because the soil organic carbon will not change greatly within a month. The correlation

415   between annual soil CO consumption and atmospheric CO concentration is 0.91 at the

416   global scale because the atmospheric CO concentration, air temperature, soil

417   temperature dominate the annual consumption rate. At monthly step, this correlation is -

418   0.48 because global atmospheric CO concentrations are high in winter and low in

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419 summer while the simulated soil CO consumption shows an opposite monthly

420 variation(Table 6, Figure 12), suggesting that other factors such as precipitation, air

421 temperature, and soil temperature are major controls for monthly CO fluxes.

422

423 **4.3. Model Uncertainties & Limitations**

424   Due to the lack of long-period observational data of CO flux and associated

425 environmental factors, the model parameterization using SCE-UA-R method can only

426 be done for 4 ecosystem types including boreal forest, temperate coniferous forest,

427 temperate deciduous forest and grassland, with RMSE ranging from 0.56 to 1.47 mg m^{-2}

428 ^{2} day^{-1}. Tropical forest calibration is only conducted using a very limited amount of lab

429 experiment data, but tropical areas are hotspots for CO soil-atmosphere exchange.

430 Besides, tropical forest SOC for top 30cm can be really high according to observations.

431 TEM model may underestimate the top 30cm SOC, which will underestimate production

432 rates, especially in tropical region. The large deviation for tropical savanna (which is

433 mosaic of tropical forest and grassland ecosystems) may be due to using outside air

434 temperature to represent inside air temperature of transparent chamber observations

435 (Varella et al., 2004), and uncertain tropical forest parameterization. We used the

436 conclusion from van Asperen et al. (2015) and only considered the thermal-degradation

437 process for CO production. Photo-degradation process and biological formation process

438 were not considered due to lacking understanding of these processes. Although we

439 focused on natural ecosystems in this study, land-use change, agriculture activity, and

440 nitrogen deposition also affect the soil CO consumption and production (King, 2002;

441 Chan & Steudler, 2006). For instance, soil CO consumption in agriculture ecosystems is

442 0 to 9 mg CO m^{-2} day^{-1} in Brazil (King & Hungria, 2002). We used grass land or forest

443 ecosystem to represent agriculture areas in CODM module. Our future study shall

444 include these processes and factors.

445

446 **5. Conclusions**

447   We analyzed the magnitude, spatial pattern, and the controlling factors of

448 atmosphere-soil CO exchange at the global scale for the 20th and 21st centuries using

449   a calibrated process-based biogeochemistry model. Major processes include

450   atmospheric CO diffusion into soils, microbial oxidation removal of CO, and CO

451   production through chemical reaction. We found that air temperature and soil

452   temperature play a dominant role in determining annual soil CO consumption and

453   production while precipitation, air temperature, and soil temperature are the major

454   controls for the monthly consumption and production. Atmospheric CO concentrations

455   will be important for annual CO consumption. We estimated that the global annual CO

456   consumption, production and net flux for the 20th century are 132-154, 29-36 and 112-

457   119 Tg CO yr^{-1}, respectively, when using a constant atmospheric CO concentration. The

458   CO consumption reaches 180-197 Tg CO yr^{-1} during 2000-2013 when using

459   atmospheric CO concentrations observed by the MOPITT satellite. Tropical evergreen

460   forest, savanna and deciduous forest areas are the largest sinks accounting for 66% of

461   the total CO consumption, while the Northern Hemisphere consumes 60% of the global

462   total. During the 21st century, the predicted net CO flux will reach 126-150 Tg CO yr^{-1} in

463   the 2090s, primarily because of increasing air temperature. The areas near the equator,

464   eastern Asia, Europe and eastern US will become the sink hotspots because they have

465   warm and moist soils. This study calls for long-period observations of CO flux for

466   various ecosystem types to improve models. The effects of land-use change, agriculture

467   activities, nitrogen deposition, photo-degradation and biological formation shall also be

468   considered to improve future quantification of soil CO fluxes.

469

470

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Table 1. Model parameterization sites for thermal and hydrology modules (site No. 1-4) and for CODM module (site No. 5-13)

No.	Site Name	Location	Vegetation	Driving Climate	Observed Data	Source and Comments
1	Poker Flat Research Range Black Spruce Forest (US_PRR)	147°29'W/65°7'N	Boreal Evergreen Needle Leaf Forests	Site Observation & ERA Interim	Soil Temperature and Moisutre of 2011-2014	Suzuki (2016)
2	Morgan Monroe State Forest (US_MMS)	86°25'W/39°19'N	Temperate Deciduous Broadleaf Forests	Site Observation & ERA Interim	Soil Temperature and Moisutre of 1999-2014	Philip and Novick (2016)
3	Santarem, Tapajos National Forest (STM_K83)	54°56'W/3°3'S	Tropical Moist Forest	Site Observation & ERA Interim	Soil Temperature and Moisutre of 2000-2004	SALESKA et al. (2013)
4	Bananal Island Site (TOC_BAN)	50°08'W/9°49'S	Tropical Forest-Savanna	Site Observation & ERA Interim	Soil Temperature and Moisutre of 2003-2006	SALESKA et al. (2013)
5	Eastern Finland (EF)	27°14'E/63°9'N	Boreal Grassland	Site Observation & ERA Interim	CO flux of April-November,2011	Pihlatie et.al. (2016)
6	Viterbo, Italy (VI)	11°55'E/42°22'N	Mediterranean Grassland	Site Observation & ERA Interim	CO flux of August, 2013	van Asperen et al. (2015)
7	Brasilia, Brazil (BB)	47°51'W/15°56'S	Tropical Savanna	Site Observation & CRU	CO flux of October 1999 to July 2001	Varella et al. (2004)
8	Orange County, North Carolina (OC)	79°7'W/35°58'N	Temperate Coniferous Forest	AMF_US-Dk3 2002-2003	CO flux of March 2002 to March 2003	Fisher (2003)
9	Tsukuba Science City, Japan (TSC)	140°7'E/36°01'N	Temperate Mixed Forest	Site Observation & ERA Interim	CO flux of July 1996 to September 1997	Yonemura et. al. (2000)
10	Manitoba, Canada (CBS)	96°44'W/56°09'N	Boreal Pine Forest	Site Observation & AMF_CA-Man	CO flux of June-August, 1994	Kuhlbusch et. al (1998)
11	Scotland, U.K. (SUK)	3°12'W/55°51'N	Temperate Deciduous Forests	ERA Interim 1995	CO flux of 1995	Moxley and Smith (1998)
12	Alaska, USA (AUS)	147°41'W/64°52'N	Boreal wetland	CRU 1991	CO flux of Lab Experiment,1991	Funk et al. (1994)
13	Guayana Shield,Bolivar State,Venezuela (GBV)	62°57'W/7°51'N	Tropical Smideciduous Forest	CRU 1985	CO flux of Lab Experiment,1985	Scharffe et al. (1990)

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Table 2. Ecosystem-specific parameters in the CODM module[a]

	Ecosystem Type	k_{CO} (ul CO l^{-1})	V_{max} (ug CO g^{-1}h^{-1})	T_{ref} (°C)	$Q10$ (Unitless)	M_{min} ($\frac{v}{v}$)	M_{max} ($\frac{v}{v}$)	M_{opt} ($\frac{v}{v}$)	E_{SOC}	F_{SOC} ($\frac{g}{g}$)	$\frac{Ea_{ref}}{R}$ (K)	PM_{ref} ($\frac{v}{v}$)	PT_{ref} (°C)
1	Alpine Tundra & Polar Desert	36.00	0.78	4.00	1.80	0.10	1.00	0.55	3.00	0.33	7700	0.25	30.00
2	Wet Tundra	36.00	0.70	4.00	1.80	0.25	1.00	0.55	3.00	0.42	7700	0.25	30.00
3	Boreal Forest	27.34	1.18	9.81	1.60	0.15	0.64	0.53	2.98	0.50	8827	0.35	26.99
4	Temperate Coniferous Forest	42.64	2.15	6.90	1.87	0.02	0.96	0.53	2.86	0.50	8404	0.38	31.52
5	Temperate Deciduous Forest	40.16	2.43	8.54	1.51	0.17	0.81	0.51	2.45	0.50	8801	0.35	37.44
6	Grassland	42.41	0.49	11.27	1.65	0.16	0.82	0.51	3.09	0.42	14165	0.24	12.29
7	Xeric Shrublands	8.00	0.30	4.00	1.50	0.10	1.00	0.55	3.00	0.33	7700	0.25	30.00
8	Tropical Forest	45.00	2.00	4.00	1.50	0.10	1.00	0.55	3.80	0.50	14000	0.50	18.00
9	Xeric Woodland	8.00	0.30	4.00	1.50	0.10	1.00	0.55	3.00	0.50	7700	0.25	30.00
10	Temperate Evergreen Broadleaf Forest	40.16	2.43	8.54	1.51	0.17	0.81	0.51	2.45	0.50	8801	0.35	37.44
11	Mediterranean Shrubland	45.00	1.50	4.00	1.50	0.10	1.00	0.55	3.00	0.33	7700	0.25	30.00
**	Largest Potential Value	51.00	11.1	15.00	2.00	0.30	1.00	0.60	3.80	--	15000	0.60	40.00

[a] k_{CO} is the half-saturation constant for soil CO concentration; V_{max} is the specific maximum CO oxidation rate; T_{ref} is the reference temperature to account for soil temperature effects on CO consumption; $Q10$ is the an ecosystem-specific Q10 coefficient to account for soil temperature effects on CO consumption; M_{min}, M_{max}, M_{opt} are the minimum, optimum, and maximum volumetric soil moistures of oxidation reaction to account for soil moisture effects on CO consumption; E_{SOC} is an estimated nominal CO production factor, similar as Potter et al. (1996) (10^{-4} mg CO m^{-2} d^{-1} per g SOC m^{-2}); F_{SOC} is a constant fraction of top 20cm SOC compared to total amount of SOC to account for SOC effects on CO production; Ea_{ref}/R is the is the ecosystem-specific activation energy divided by gas constant to account for the reaction rate of production; PM_{ref} is the reference moisture to account for soil temperature effects on CO production; PT_{ref} is the reference temperature to account for soil temperature effects on CO production

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722 **Table 3.** Regional soil CO consumption, net flux and production(Tg CO yr^{-1}) during 1901- 2013 (E1) and during

723 2000-2013 withMOPITT data transient CO surface concentration

	South-45°S		45°S-0°		0°-45°N		45°N-North		Global	
	E1[a]	E2[b]	E1	E2	E1	E2	E1	E2	E1	E2
Consumption	-0.20	-0.22	-58.36	-75.77	-64.78	-91.66	-17.24	-18.90	-140.58	-186.55
Net flux	-0.12	-0.13	-43.39	-59.34	-51.58	-77.17	-13.27	-14.63	-108.35	-151.27
Production	0.09	0.09	14.98	16.43	13.20	14.49	3.97	4.27	32.23	35.28

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725 [a]E1 represents the simulation with constant CO surface concentration data;

726 [b]E2 represents the simulation with MOPITT transient CO surface concentration data.

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747 **Table 4.** Global soil CO consumption, net flux and production in different ecosystems during 1901-2013

Vegetation Type	Area (10⁶ km²)	Pixels	Consumption (Tg CO yr⁻¹)	Net flux (Tg CO yr⁻¹)	Production (Tg CO yr⁻¹)
Alpine Tundra & Polar Desert	5.28	3580	-0.95	-0.73	0.22
Wet Tundra	5.24	4212	-1.06	-0.52	0.54
Boreal Forest	12.47	7578	-7.19	-5.53	1.66
Forested Boreal Wetland	0.23	130	-0.13	-0.09	0.04
Boreal Woodland	6.48	4545	-2.40	-1.53	0.88
Non-Forested Boreal Wetland	0.83	623	-0.33	-0.17	0.16
Mixed Temperate Forest	5.25	2320	-6.31	-5.82	0.49
Temperate Coniferous Forest	2.49	1127	-2.78	-2.49	0.29
Temperate Deciduous Forests	3.65	1666	-3.26	-3.02	0.24
Temperate Forested Wetland	0.15	60	-0.22	-0.21	0.01
Tall Grassland	3.63	1567	-1.28	-0.37	0.91
Short Grassland	4.71	2072	-0.93	-0.25	0.68
Tropical Savanna	13.85	4666	-15.83	-10.41	5.42
Xeric Shrubland	14.71	5784	-1.61	-1.33	0.28
Tropical Evergreen Forest	17.77	5855	-66.12	-51.28	14.84
Tropical Forested Wetland	0.55	178	-2.51	-2.05	0.46
Tropical Deciduous Forest	4.69	1606	-11.20	-8.48	2.72
Xeric Woodland	6.85	2387	-6.53	-5.57	0.96
Tropical Forested Floodplain	0.15	50	-0.64	-0.54	0.11
Desert	11.61	4170	-0.49	-0.45	0.05
Tropical Non-forested Wetland	0.06	19	-0.02	-0.01	0.01
Tropical Non-forested Floodplain	0.36	120	-0.20	-0.11	0.09
Temperate Non-Forested Weland	0.34	120	-0.22	-0.10	0.13
Temperate Forested Floodplain	0.10	48	-0.10	-0.10	0.00
Temperate Non-forested Floodplain	0.10	45	-0.04	-0.02	0.01
Wet Savanna	0.16	59	-0.27	-0.21	0.06
Salt Marsh	0.09	35	-0.04	-0.01	0.02
Mangroves	0.12	38	-0.39	-0.32	0.07
Temperate Savannas	6.83	2921	-3.17	-2.63	0.54
Temperate Evergreen Broadleaf	3.33	1268	-3.60	-3.39	0.21
Mediterranean Shrubland	1.47	575	-0.75	-0.60	0.15
Total	133.56	59424	-140.58	-108.35	32.23

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754 **Table 5.** Sensitivity of global CO consumption, net flux and production (units are Tg CO yr^{-1}) to changes in

755 atmospheric CO, soil organic carbon (SOC), precipitation (Prec) and air temperature (AT)

	Baseline	CO +30%	CO -30%	SOC +30%	SOC -30%	Prec +30%	Prec -30%	AT +3°C	AT -3°C
Consumption	-147.65	-164.14	-131.12	-175.37	-119.90	-150.72	-143.50	-190.59	-114.83
Change (%)	0.00	-11.17	11.19	-18.78	18.79	-2.08	2.81	-29.09	22.23
Net flux	-113.65	-130.15	-97.12	-131.18	-96.10	-116.97	-109.32	-144.23	-89.58
Change (%)	0.00	-14.51	14.54	-15.42	15.44	-2.92	3.81	-26.90	21.18
Production	33.99	33.99	33.99	44.19	23.80	33.74	34.17	46.36	25.25
Change (%)	0.00	0.00	0.00	30.00	-30.00	-0.75	0.53	36.39	-25.72

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777 **Table 6.** Effects of annual and monthly climate precipitation (Prec), air temperature (Tair), soil organic carbon

778 (SOC), soil temperature (Tsoil), soil moisture (Msoil) and atmospheric CO (CO air) on absolute values of

779 consumption, production and net flux for different regions and the globe during the 20th Century

		Monthly					Annual				
		North-45°N	45°N-0°	0°-45°S	45°S-South	Global	North-45°N	45°N-0°	0°-45°S	45°S-South	Global
Prec	Consumption	0.91	0.96	0.92	-0.34	0.87	0.65	0.21	0.26	0.13	0.52
	Production	0.91	0.70	0.45	-0.34	0.82	0.63	0.10	0.15	-0.11	0.47
	Net flux	0.91	0.97	0.94	-0.33	0.87	0.65	0.25	0.31	0.32	0.54
Tair	Consumption	0.97	0.98	0.91	0.96	0.95	0.92	0.93	0.88	0.84	0.91
	Production	0.96	0.83	0.72	0.98	0.94	0.92	0.92	0.91	0.95	0.91
	Net Flux	0.97	0.97	0.88	0.90	0.95	0.91	0.92	0.85	0.62	0.91
SOC	Consumption	-0.19	0.07	0.21	-0.01	0.15	0.68	0.90	0.92	0.47	0.92
	Production	-0.19	0.31	0.47	-0.02	0.24	0.72	0.92	0.92	0.50	0.93
	Net Flux	-0.19	0.03	0.14	0.00	0.13	0.67	0.88	0.91	0.38	0.91
Tsoil	Consumption	0.97	0.98	0.92	0.96	0.95	0.94	0.93	0.88	0.85	0.95
	Production	0.97	0.83	0.72	0.98	0.94	0.94	0.92	0.91	0.96	0.95
	Net Flux	0.98	0.97	0.88	0.90	0.95	0.93	0.93	0.86	0.63	0.95
Msoil	Consumption	0.85	0.96	0.92	0.19	0.76	0.03	0.22	0.14	0.26	0.22
	Production	0.85	0.75	0.44	0.14	0.69	-0.02	0.12	0.02	0.05	0.17
	Net Flux	0.84	0.96	0.95	0.25	0.77	0.04	0.26	0.19	0.40	0.24
CO Air	Consumption	-0.66	-0.76	-0.29	0.14	-0.48	0.87	0.88	0.81	0.98	0.91
	Production	-0.70	-0.66	0.08	-0.40	-0.66	-0.36	-0.48	-0.54	-0.44	-0.57
	Net Flux	-0.64	-0.73	-0.35	0.55	-0.41	0.92	0.91	0.88	0.99	0.94

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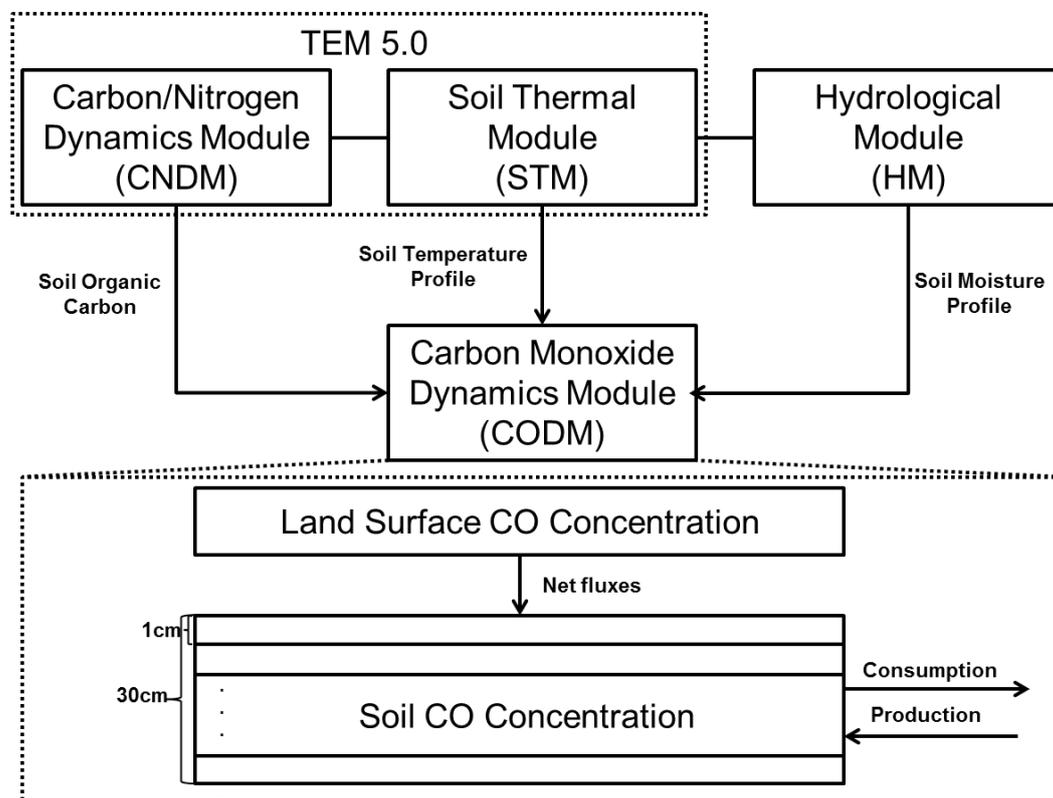
Figure 1. The model framework includes a carbon and nitrogen dynamics module (CNDM), a soil thermal module (STM) from Terrestrial Ecosystem Model (TEM) 5.0 (Zhuang et al., 2001, 2003), a hydrological module (HM) based on a Land Surface Module (Bonan, 1996; Zhuang et al., 2004), and a carbon monoxide dynamics module (CODM). The detailed structure of CODM includes land surface CO concentration as top boundary and thirty 1 cm thick layers (totally 30 cm) where consumption and production would happen inside.
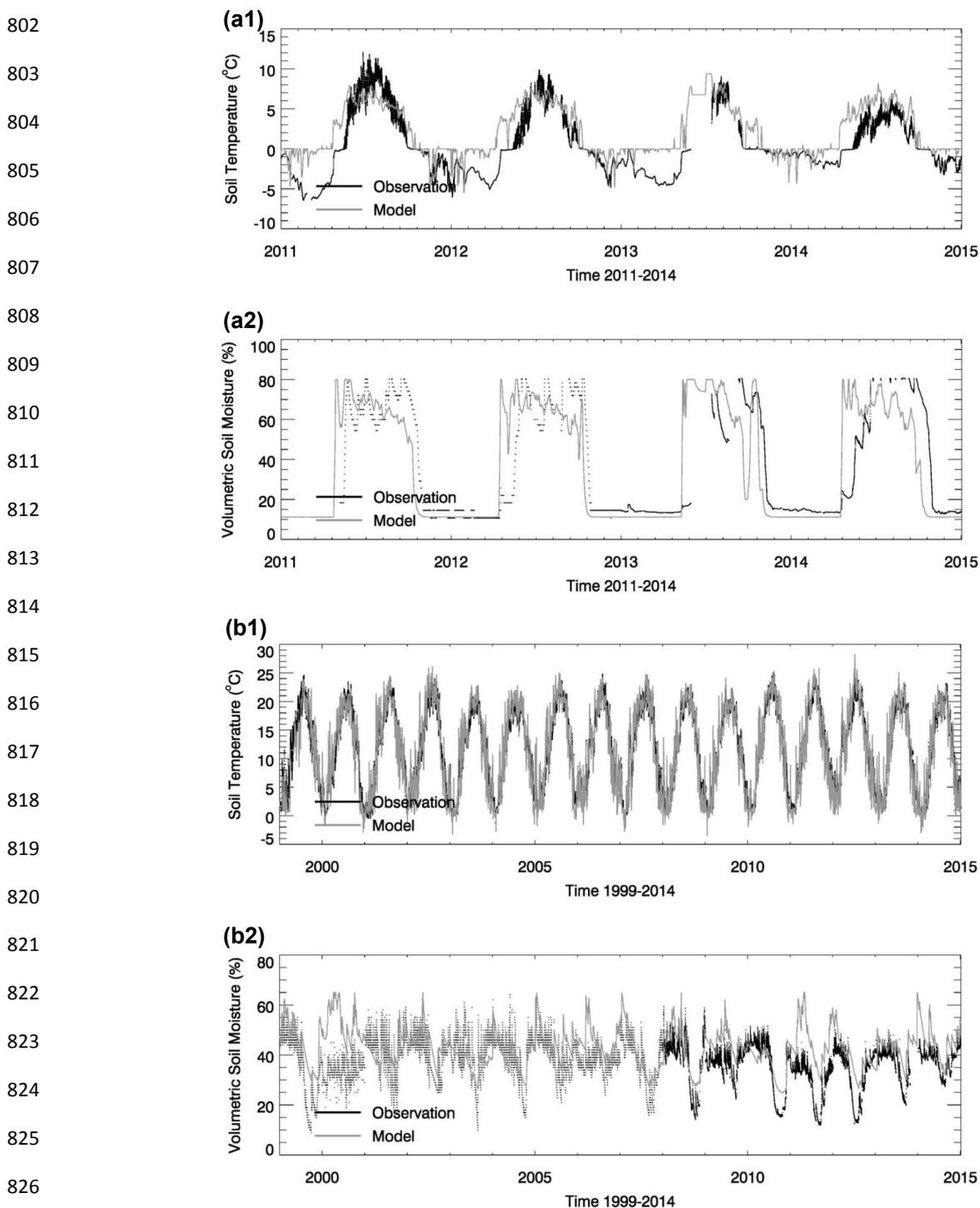
Figure 2. Evaluation of thermal and hydrology module at four sites: (a) Boreal Evergreen Needle Leaf Forests, (b) Temperate Deciduous Broadleaf Forests. (1) shows the soil temperature comparison between model simulations (gray line) and observations (black line) and (2) shows the soil moisture comparison between model simulations (gray line) and observations (black line).
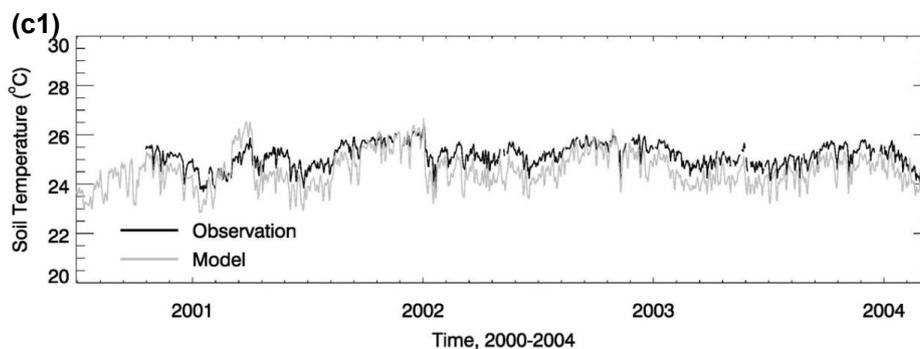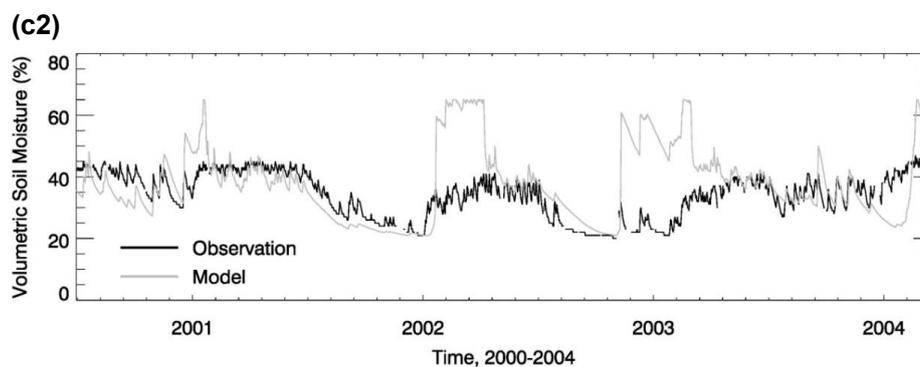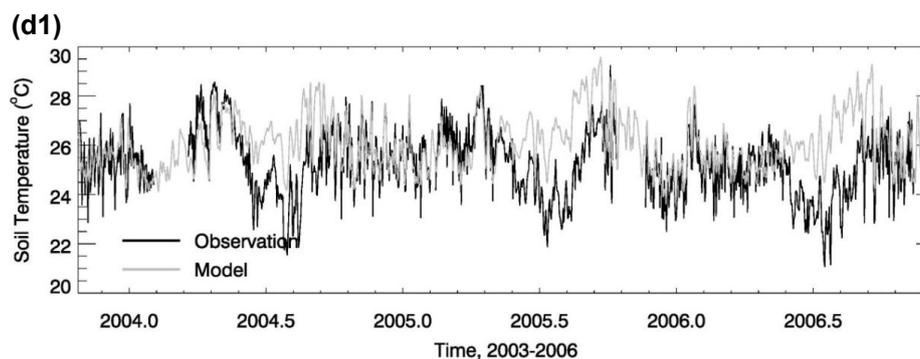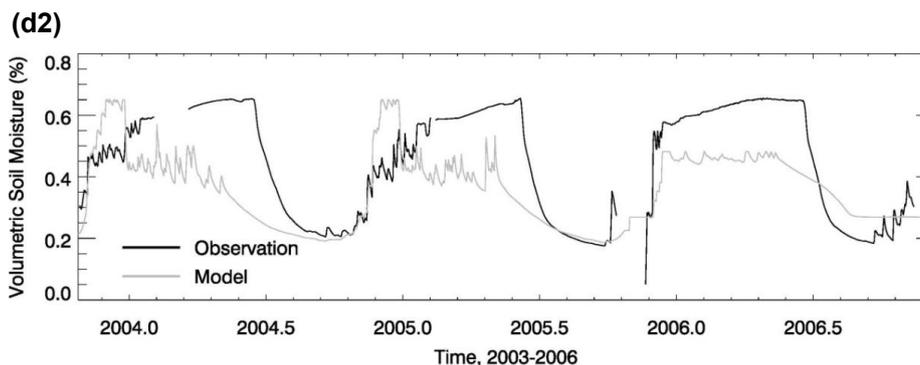
Figure 2. Contd. Evaluation of thermal and hydrology module at four sites: (c) Tropical Moist Forest, (d) Tropical Forest-Savanna. (1) shows the soil temperature comparison between model simulations (gray line) and observations (black line) and (2) shows the soil moisture comparison between model simulations (gray line) and observations (black line)

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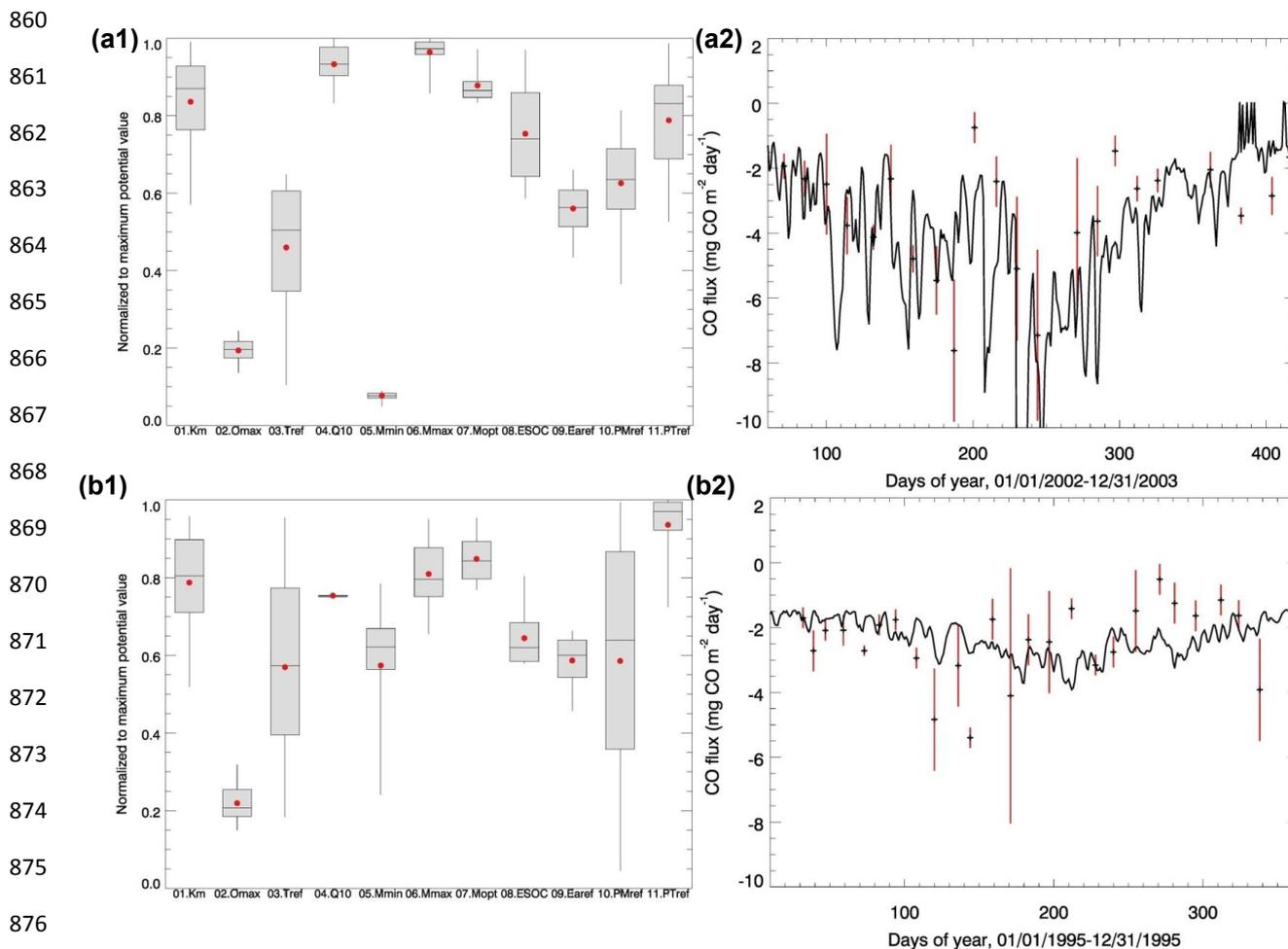
Figure 3. Parameter ensemble experiment results: Each parameter has 50 calibrated values generated from running SCE-UA-R 50 times independently. Parameters are normalized to their largest potential values described in Table 2. (a1) and (a2) are temperate coniferous forest normalized parameter distribution boxplots and CO flux comparisons between model simulations (solid line, using mean value of parameters) and observations ("+", red lines represent error bar), respectively. For each box, line top, box top, horizontal line inside box, box bottom and line bottom represent maximum, third quartile, median, first quartile and minimum of 50 parameter values. Red dot represents the mean value of 50 parameter values. (b1) and (b2) are plots for temperate deciduous forest; (c1) and (c2) are for boreal forest; (d1) and (d2) are for grassland. Grassland observation data is the sum of hourly observations so there is no error bar presented.

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Figure 3. Contd. Parameter ensemble experiment results: Each parameter has 50 calibrated values generated from running SCE-UA-R 50 times independently. Parameters are normalized to their largest potential values described in Table 2. (a1) and (a2) are temperate coniferous forest normalized parameter distribution boxplots and CO flux comparisons between model simulations (solid line, using mean value of parameters) and observations ("+", red lines represent error bar), respectively. For each box, line top, box top, horizontal line inside box, box bottom and line bottom represent maximum, third quartile, median, first quartile and minimum of 50 parameter values. Red dot represents the mean value of 50 parameter values. (b1) and (b2) are plots for temperate deciduous forest; (c1) and (c2) are for boreal forest; (d1) and (d2) are for grassland. Grassland observation data is the sum of hourly observations so there is no error bar presented.

Figure 4. Historical global land surface mean climate, and simulated global mean soil moisture, soil temperature and SOC for the period 1901-2013.
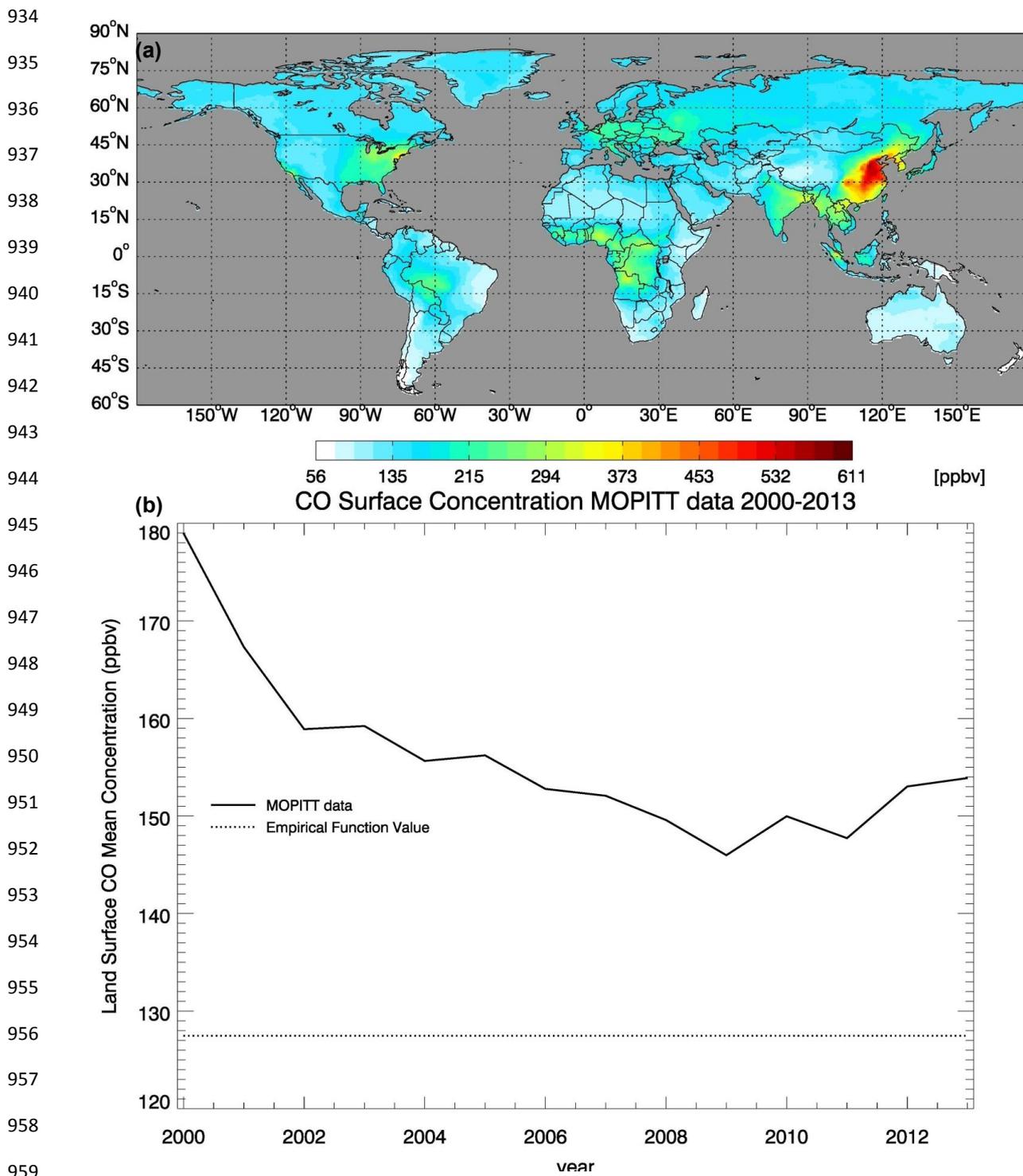
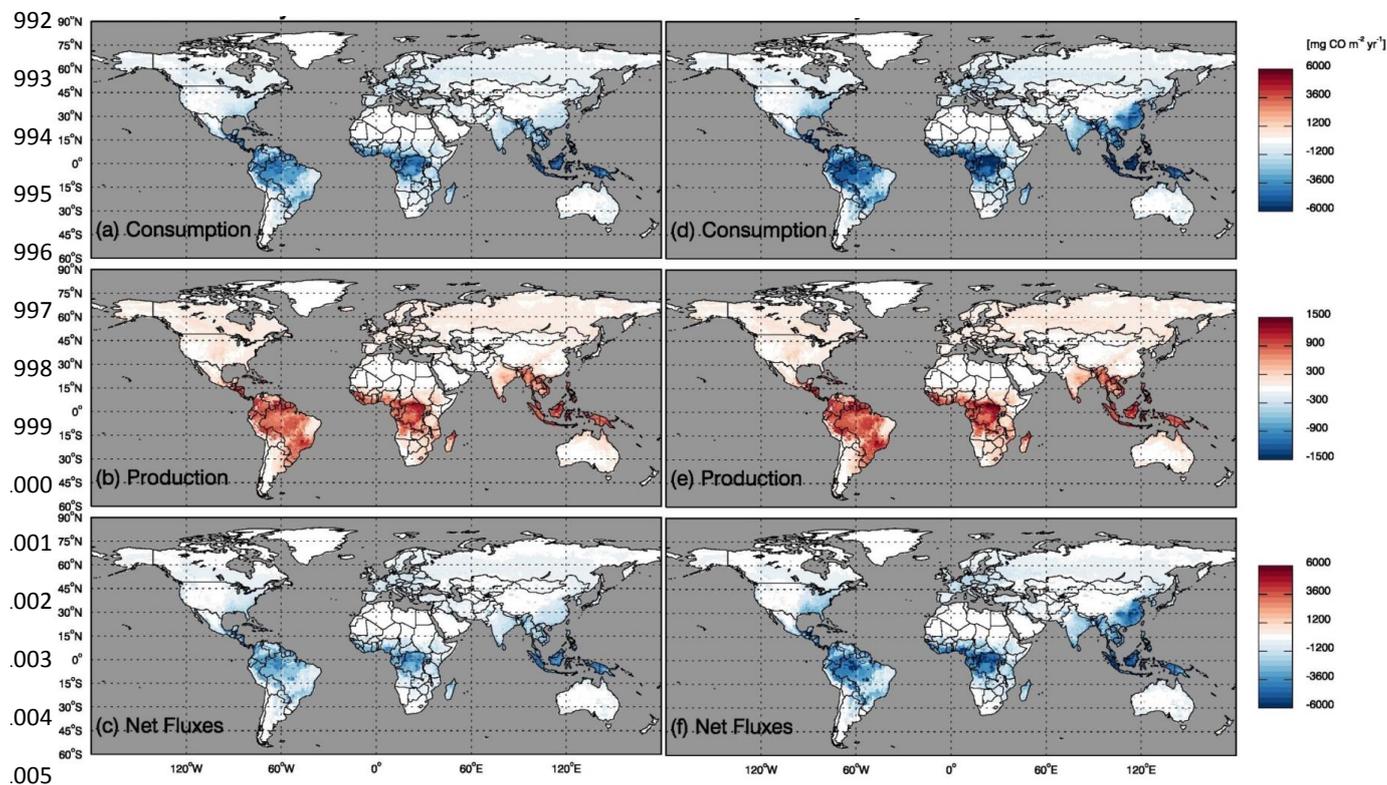**Figure 5.** CO surface concentration data from MOPITT satellite (ppbv): (a) global mean CO surface concentrations from MOPITT during 2000-2013; (b) the CO annual surface concentrations from both MOPITT and empirical functions (Potter et al., 1996).
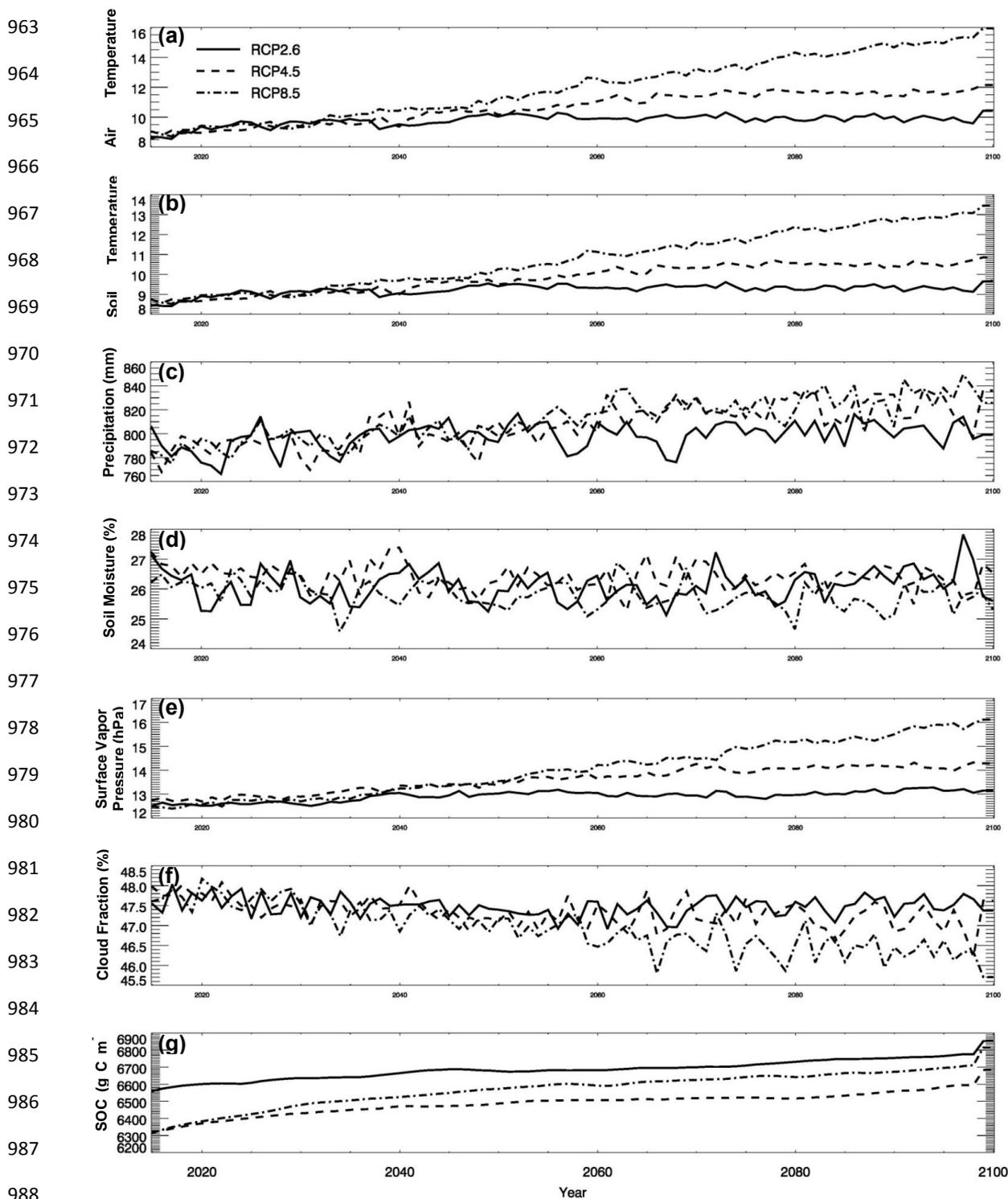
Figure 6. Global mean climate from RCP2.6, RCP4.5 and RCP8.5 data sets and simulated global mean soil temperature, moisture and SOC: (a)-(g) are land surface air temperature (°C), soil temperature (°C), precipitation (mm), soil moisture (%), surface water vapor pressure (hpa), cloud fraction (%), and SOC (mg m^{-2}), respectively.

Figure 7. Global annual mean soil CO fluxes (mg CO m^{-2} yr^{-1}) during 1901-2013, estimated using constant CO concentration data (left side) and mean annual global soil CO fluxes during 2000-2013 using MOPITT CO atmospheric surface concentration data (right side)
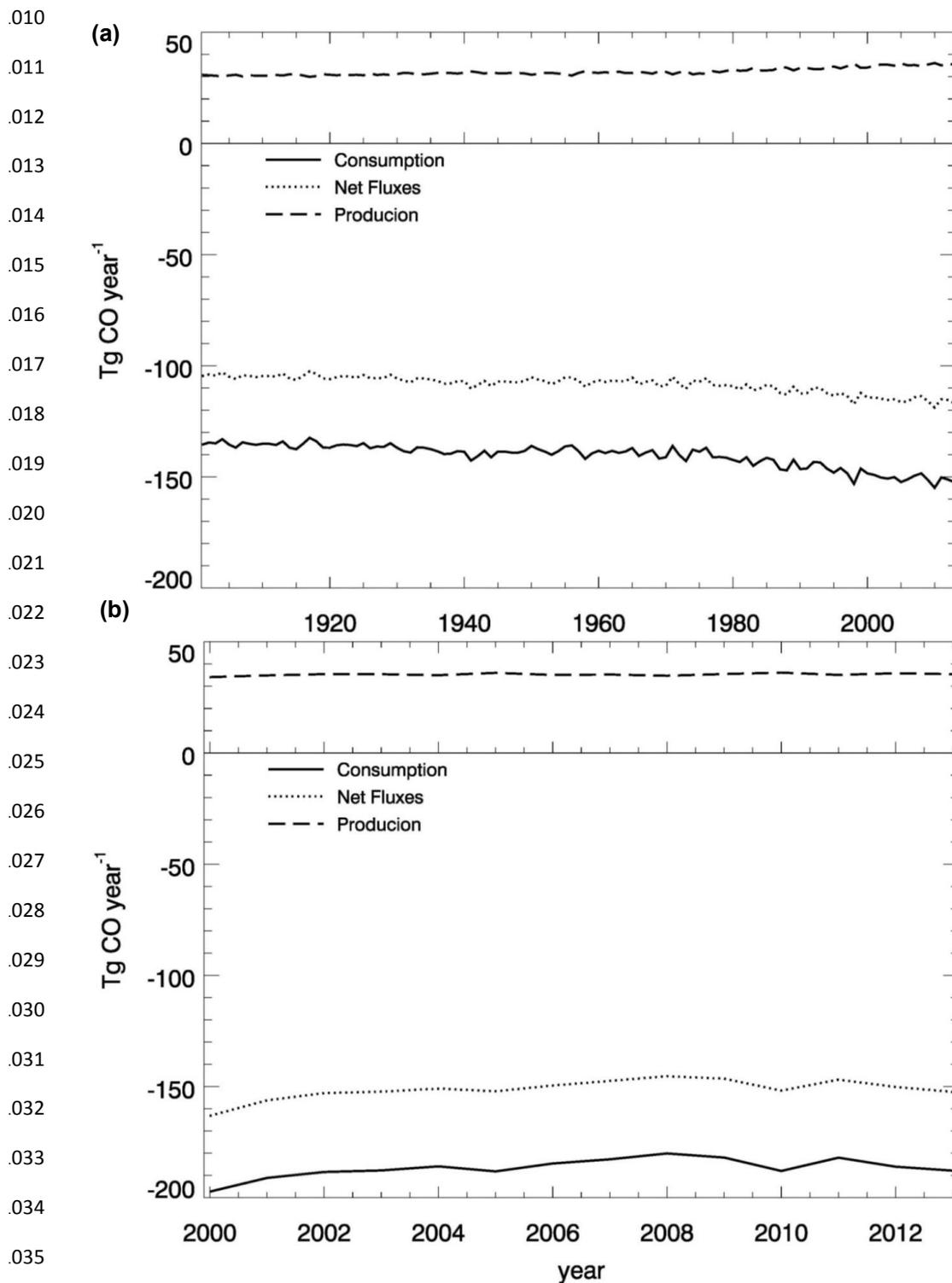
Figure 8. Global mean CO consumption, production and net flux(Tg CO yr^{-1}): (a) from 1901 to 2013, estimated with constant CO surface concentration data and (b) from 2000 to 2013 with MOPITT CO surface concentration data.
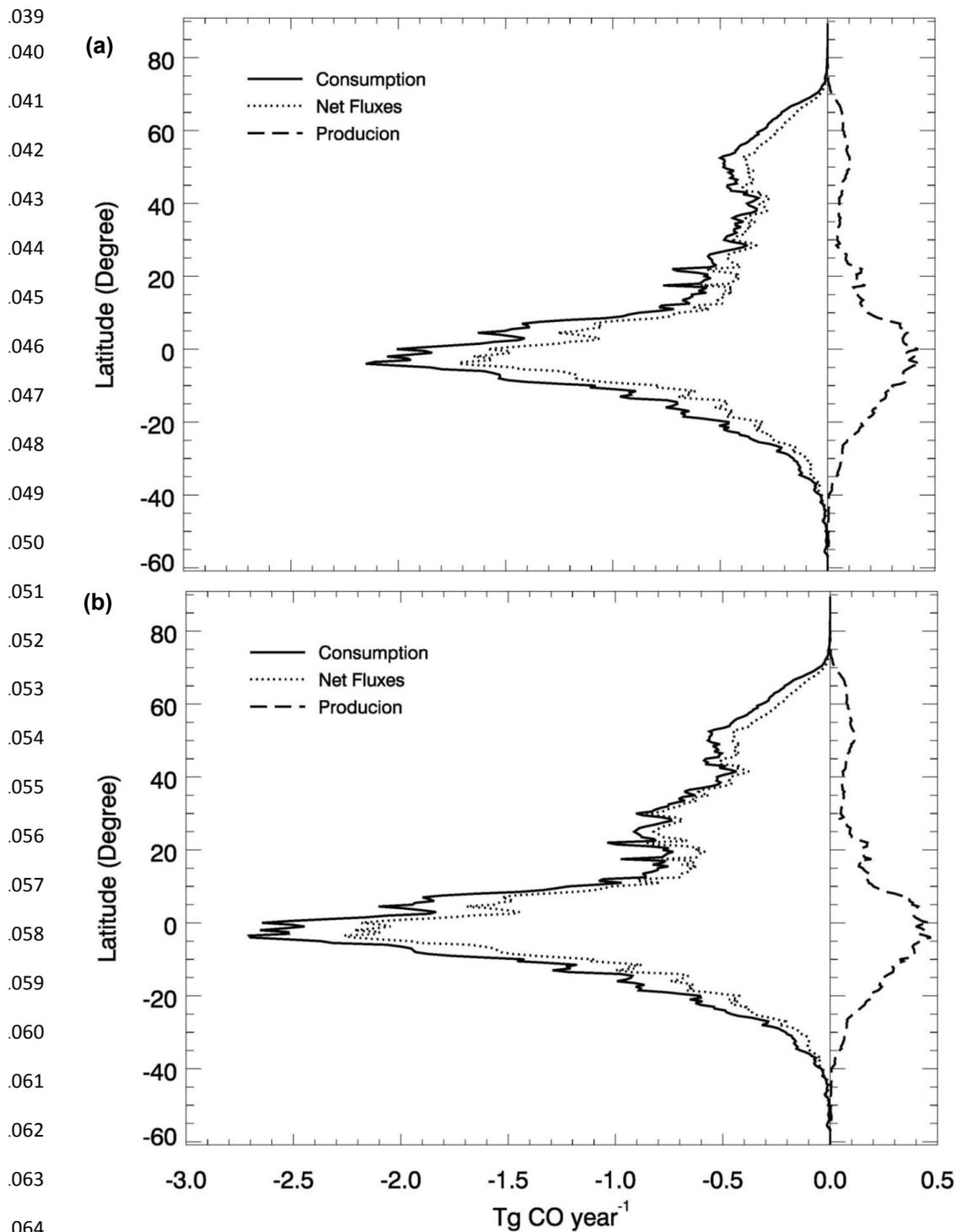
Figure 9. Global annual mean latitudinal distributions of soil CO consumption, production and net flux: (a) during 1901-2013 (Tg CO yr^{-1}) estimated with constant CO surface concentration data and (b) during 2000-2013 estimated with MOPITT CO surface concentration data.
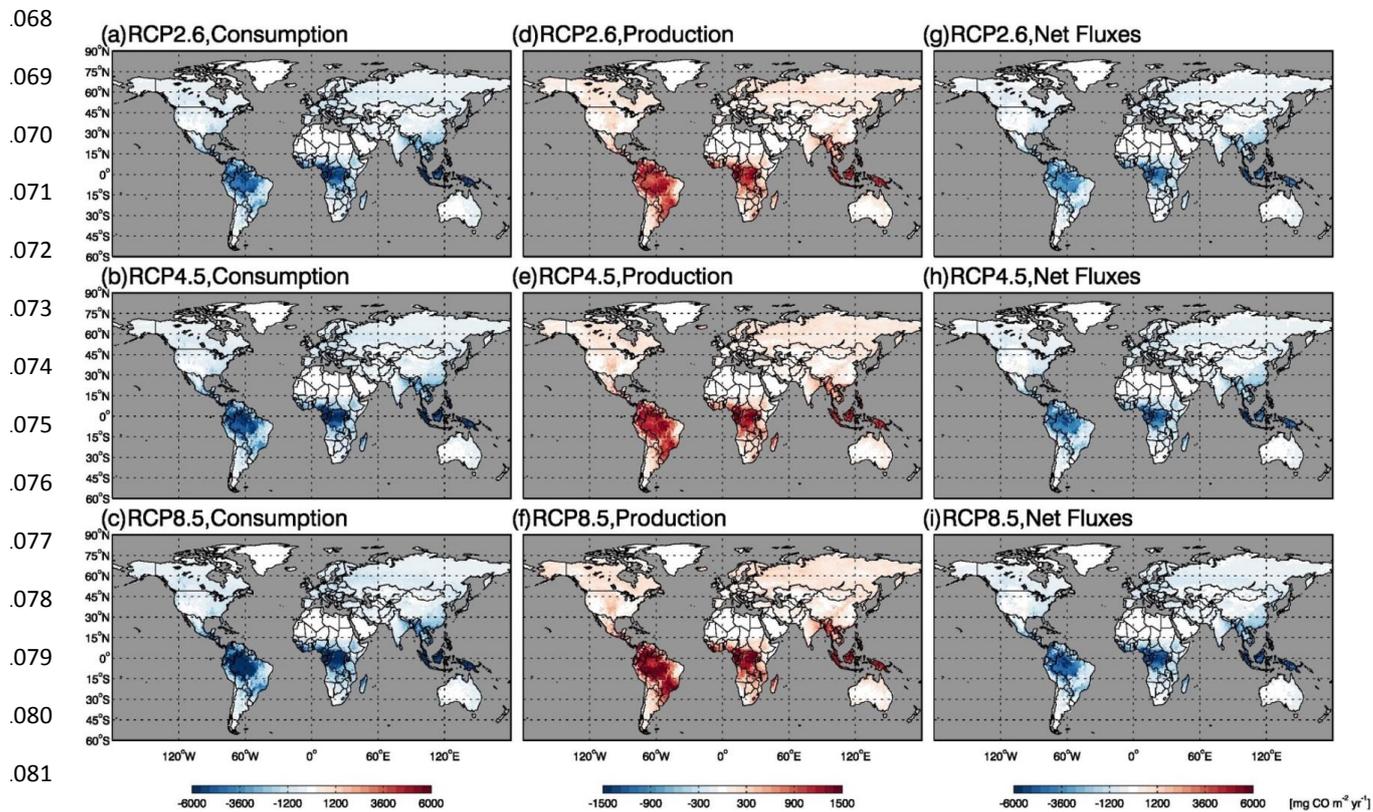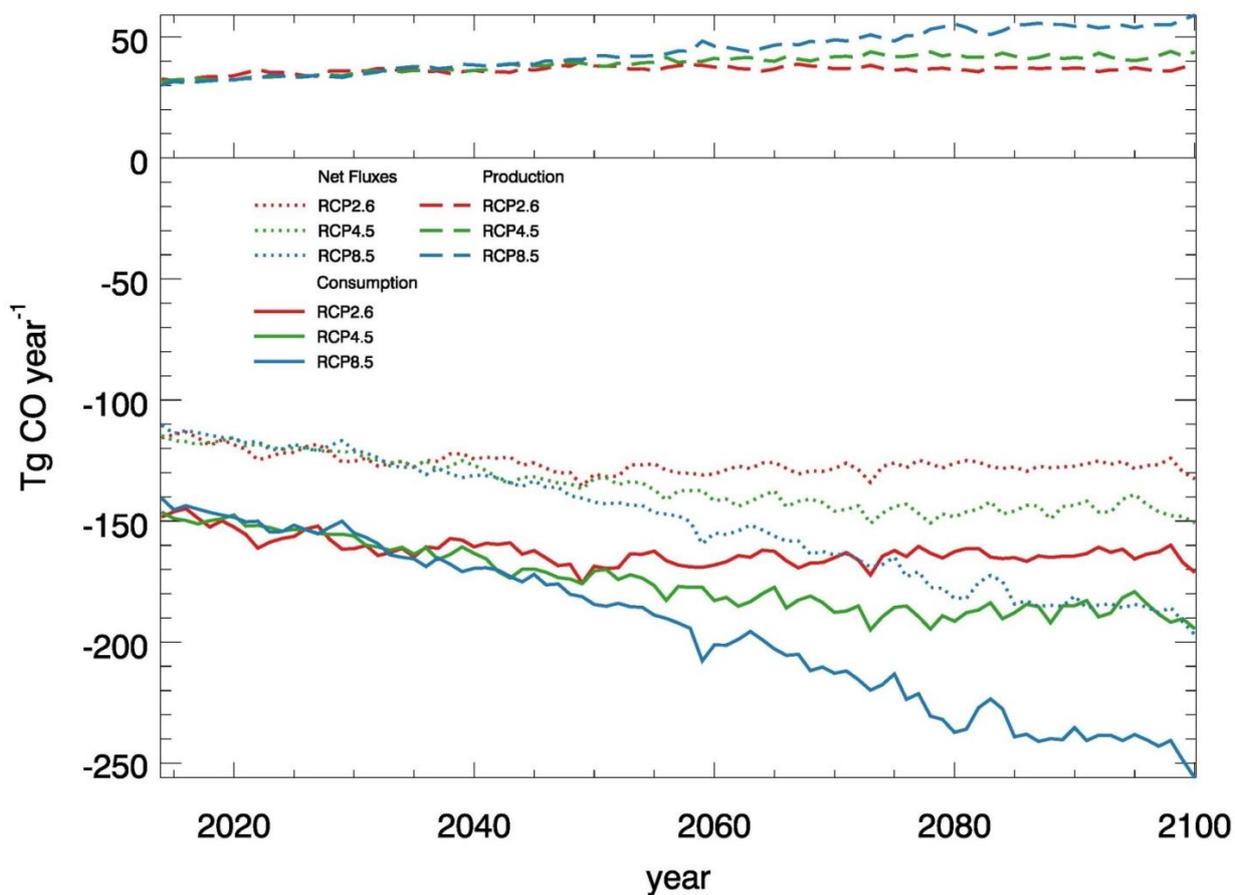
Figure 10. Global annual mean CO consumption, production and net flux (mg CO m^{-2} yr^{-1}) under future climate scenarios RCP2.6, RCP4.5 and RCP8.5 during 2014-2100
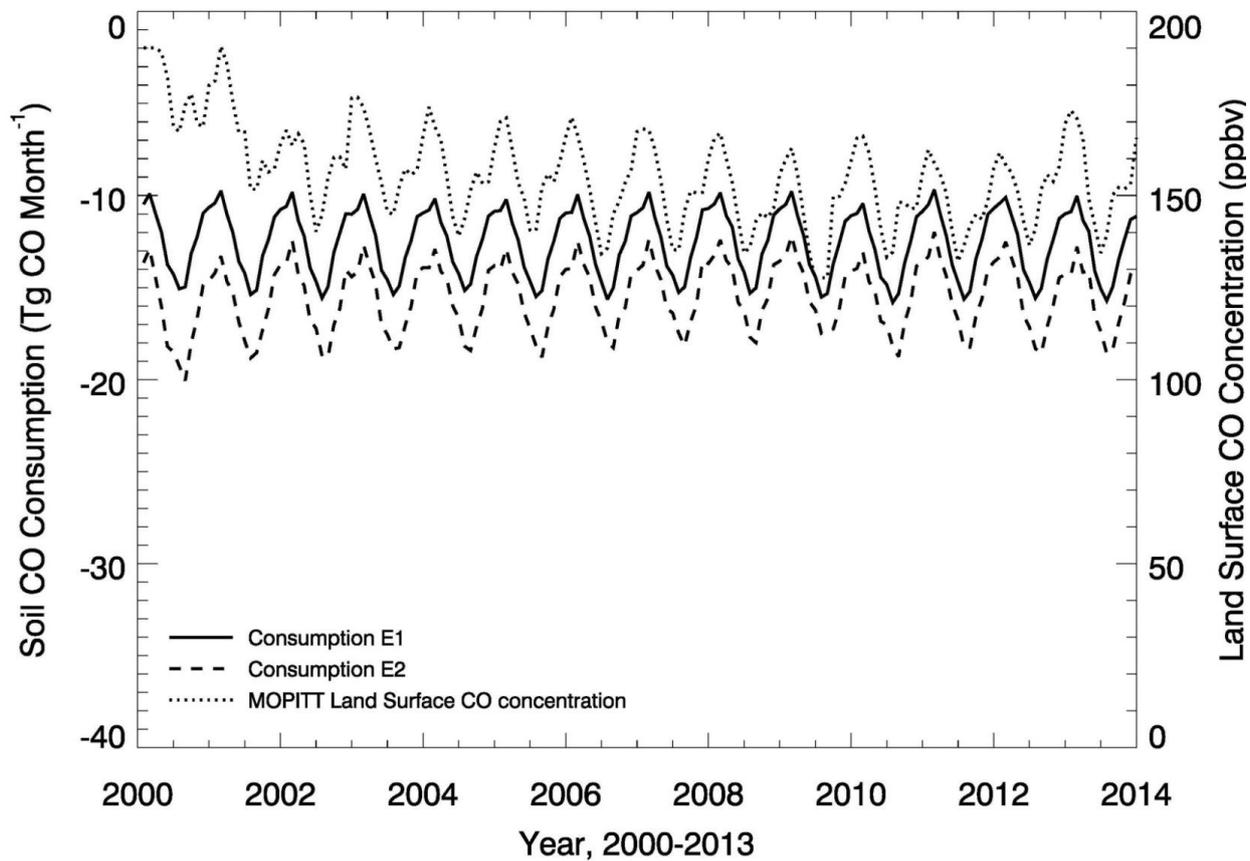
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Figure 11. Future Global mean soil CO consumption, net flux and production (Tg CO yr^{-1}) under future climate scenarios RCP2.6, RCP4.5 and RCP8.5 during 2014-2100

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.090 simulations E1 and E2 (Tg CO mon^{-1})