



Drivers of trends in
land carbon fluxes
and BVOC emissions

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Distinguishing the drivers of trends in land carbon fluxes and plant volatile emissions over the past three decades

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Abstract

The terrestrial biosphere has experienced dramatic changes in recent decades. Estimates of historical trends in land carbon fluxes remain uncertain because long-term observations are limited on the global scale. Here, we use the Yale Interactive terrestrial Biosphere (YIBs) model to estimate decadal trends in land carbon fluxes and emissions of biogenic volatile organic compounds (BVOCs) and to identify the key drivers for these changes during 1982–2011. Driven with hourly meteorology from WFDEI (WATCH Forcing Data methodology applied to ERA-Interim data), the model simulates an increasing trend of 297 TgCa^{-2} in gross primary productivity (GPP) and 185 TgCa^{-2} in the net primary productivity (NPP). CO_2 fertilization is the main driver for the flux changes in forest ecosystems, while meteorology dominates the changes in grasslands and shrublands. Warming boosts summer GPP and NPP at high latitudes, while drought dampens carbon uptake in tropical regions. North of 30° N , increasing temperatures induce a substantial extension of 0.22 day a^{-1} for the growing season; however, this phenological change alone does not promote regional carbon uptake and BVOC emissions. Nevertheless, increases of LAI at peak season accounts for $\sim 25\%$ of the trends in GPP and isoprene emissions at the northern lands. The net land sink shows statistically insignificant increases of only 3 TgCa^{-2} globally because of simultaneous increases in soil respiration. In contrast, driven with alternative meteorology from MERRA (Modern Era-Retrospective Analysis), the model predicts significant increases of 59 TgCa^{-2} in the land sink due to strengthened uptake in the Amazon. Global BVOC emissions are calculated using two schemes. With the photosynthesis-dependent scheme, the model predicts increases of 0.4 TgCa^{-2} in isoprene emissions, which are mainly attributed to warming trends because CO_2 fertilization and inhibition effects offset each other. Using the MEGAN (Model of Emissions of Gases and Aerosols from Nature) scheme, the YIBs model simulates global reductions of 1.1 TgCa^{-2} in isoprene and 0.04 TgCa^{-2} in monoterpane emissions in response to the CO_2 inhibition effects. Land use change shows limited impacts on global carbon

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fluxes and BVOC emissions, but there are regional contrasting impacts over Europe (afforestation) and China (deforestation).

1 Introduction

The terrestrial biosphere interacts with the atmosphere through photosynthesis and biogenic volatile organic compound (BVOC) emissions. Annually, terrestrial ecosystems assimilate ~ 120 petagrams of carbon (Pg C) from the atmosphere (Beer et al., 2010), most of which reenters atmosphere through respiration and decomposition, resulting in a net global land carbon sink of $2.6 \pm 0.7 \text{ PgC a}^{-1}$ (Le Quere et al., 2009; Sitch et al., 2015). Global BVOC emissions are estimated to be about 1 PgC per year (Carslaw et al., 2010). These emissions are important precursors of atmospheric oxidants and aerosols, both of which affect surface air quality and exert additional regional and global chemical climate forcings (Scott et al., 2014; Unger, 2014). Observations and simulations have shown significant changes in terrestrial carbon assimilation and BVOC emissions in the past 2–3 decades (Lathiere et al., 2006; Sarmiento et al., 2010; Sindelarova et al., 2014; Sitch et al., 2015). Understanding drivers of these trends is important for the projections of future carbon fluxes, water cycle, air quality, and climatic responses.

Trends in land carbon assimilation and BVOC emissions are related to the changes in atmospheric CO_2 , meteorology, and human land use land cover change perturbations. Elevated CO_2 promotes plant photosynthesis (Ainsworth and Long, 2005) but can directly inhibit isoprene productions (Arneth et al., 2007). Warming accelerates both carbon uptake and BVOC emissions when temperature is not above the thermal optimum ($25\text{--}30^\circ\text{C}$ for photosynthesis and $35\text{--}40^\circ\text{C}$ for isoprene emission) for ecosystems that are not water-stressed (Farquhar et al., 1980; Guenther et al., 1993; Piao et al., 2013). Additional warming above thermal optimum may decrease photosynthesis but still promote respiration, reducing net carbon uptake by plants (Liang et al., 2013). Increased temperatures also indirectly influence carbon exchange and BVOC emissions through

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the extension of growing season (Piao et al., 2007). Drought decreases gross primary productivity (GPP) and net primary productivity (NPP) (Zhao and Running, 2010), but may temporally enhance isoprene emissions (Monson et al., 2007). Land use change affects the regional carbon budget and BVOC emissions through either additional emissions or land cover changes due to deforestation, forest management, and agricultural activities (Lathiere et al., 2006; Houghton, 2010).

Estimates of recent decadal global trends in the land carbon budget and BVOC emissions are limited and uncertain due to the lack of observations. The earliest site-level measurements of land carbon fluxes were set up in the 1990s (Wofsy et al., 1993). The flux tower data sets provide long-term records of regional carbon exchange with high precision but low spatial representation. In contrast, satellite products, such as GPP and NPP retrievals from the Moderate Resolution Imaging Spectroradiometer (MODIS) (Zhao et al., 2005) and isoprene emissions based on tropospheric formaldehyde columns from the Global Ozone Monitoring Experiment (Palmer et al., 2006), improve the spatial coverage but usually are available for only a relatively short time period (months to several years) and suffer from systematic biases when compared with ground measurements (e.g., Heinsch et al., 2006; Marais et al., 2012). Terrestrial biosphere models, evaluated with both site-level and satellite-based observations, are useful tools to estimate trends and attribute drivers of changes in land carbon fluxes and BVOC emissions (e.g., Mao et al., 2013; Stavrakou et al., 2014; Sitch et al., 2015).

In this study, we use the well-validated Yale Interactive Terrestrial Biosphere Model (YIBs, Yue and Unger, 2015) driven with long-term reanalysis meteorology to study the global trends of land carbon fluxes and BVOC emissions over the past three decades. We compare the simulated trends with satellite observations where available (Jeong et al., 2011; Zhu et al., 2013) and multi-model ensemble estimates (e.g., Piao et al., 2013; Sitch et al., 2015). The major goals of this study are to quantify: (1) the spatial distribution of global trends in carbon fluxes and BVOC emissions during 1982–2011, (2) the dominant drivers of these trends from elevated CO₂, changes in meteorology (temperature, radiation, and soil moisture), and human land use change, (3) the feed-

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back of biosphere, including changes in phenology and leaf area index (LAI), to the land carbon uptakes and BVOC emissions; and (4) the discrepancies in BVOC trends due to application of different isoprene emission schemes. To assess uncertainties due to the meteorological forcings (Poulter et al., 2011), we apply two independent reanalysis datasets as input and compare the corresponding model results. This study does not assess wildfire emissions because no significant global trends have been predicted with the state-of-art terrestrial models (Sitch et al., 2015). In the next section, we describe the physical processes in the model and the simulations performed for trend driver attribution. Section 3 presents the simulated global distribution and associated drivers of the changes in land carbon fluxes and BVOC emissions during the past three decades. Section 4 discusses the observational uncertainties and the model sensitivities. The last section summarizes the derived trends.

2 Data and methods

2.1 Observations

We use long-term global measurements of LAI, GPP, and NPP to validate the simulated trends. The LAI dataset for 1982–2011 is retrieved based on the Normalized Difference Vegetation Index (NDVI) from Global Inventory Modeling and Mapping Studies (GIMMS) with $1/12^\circ$ resolution and a 15 day interval (Zhu et al., 2013). We also use LAI data for 2000–2011 from the MODIS (<http://modis.gsfc.nasa.gov/>). GPP datasets of 1982–2011 are upscaled from the FLUXNET eddy covariance measurements using a biosphere model (Jung et al., 2009). As a comparison, we also use the GPP and NPP datasets for 2000–2011 from the MODIS, which have been developed based on remote sensing of biome parameters and assimilated meteorology (Zhao et al., 2005). All the datasets are interpolated to the monthly interval at the $1^\circ \times 1^\circ$ off-line YIBs model resolution.

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photosynthesis-based (PS_BVOC) isoprene scheme calculates emissions based on the electron transport-limited photosynthesis rate, canopy temperature, and intercellular CO_2 concentrations (Arneeth et al., 2007; Unger et al., 2013). The Model of Emissions of Gases and Aerosols from Nature (MEGAN) scheme applies commonly used leaf-level empirical functions of light and canopy temperature. Both schemes implement CO_2 inhibition effects on BVOC emissions parameterized as a reciprocal empirical function of intercellular $[\text{CO}_2]$ following the observations from Possell et al. (2005). For monoterpene emissions, the YIBs model applies the same temperature-dependent scheme as Lathiere et al. (2006) but with CO_2 -inhibition effects. The leaf-level BVOC emissions are integrated over the multiple canopy layers following the same approach as GPP to obtain the total canopy-level emissions.

YIBs can be used in three different configurations with increasing complexity: (1) off-line local site level, which is driven with hourly measurements of CO_2 concentrations and meteorology at flux tower sites, (2) off-line global forced with spatially uniform but annually updated CO_2 concentrations and hourly gridded reanalysis meteorology, (3) on-line coupled to the NASA ModelE2 driven with simulated meteorology by the GCM every half hour. All of the three configurations have been fully validated with carbon fluxes from 145 flux tower sites and multiple satellite products (Yue and Unger, 2015). In this study, we use the (2) off-line global version of the model, which is driven with global meteorology reanalysis data and observed CO_2 concentrations.

2.3 Simulations

We apply observed historical atmospheric CO_2 concentrations from the fifth assessment report (AR5) of the Intergovernmental Panel on Climate Change (IPCC) (Meinshausen et al., 2011). We apply an annually-varying historical transient land cover dataset (Oleson et al., 2013), which is developed based on a combination of remote sensing data from both MODIS (Hansen et al., 2003) and the Advanced Very High Resolution Radiometer (AVHRR) (Defries et al., 2000), and with land use change from Hurtt et al. (2011). We use hourly meteorological variables for 1980–2011 from two

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independent reanalysis datasets as a comparison. The first is WATCH Forcing Data methodology applied to ERA-Interim data (WFDEI, Weedon et al., 2014). The WFDEI reanalysis is an update of the WATCH Forcing Data (WFD), which is developed based on the European Centre for Medium-range Weather Forecasts (ECMWF) ERA-40 reanalysis (Uppala et al., 2005). The other is the Global Modeling and Assimilation Office (GMAO) Modern Era-Retrospective Analysis (MERRA) data set, which is generated with the GEOS-5 atmospheric general circulation model (AGCM) by assimilating both in situ and remote sensing observations (Rienecker et al., 2011). The MERRA-land product is a supplemental and improved set of land surface hydrological fields for MERRA (Reichle et al., 2011). Meteorological variables applied include surface air temperature, specific humidity, wind speed, surface pressure, total PAR, and soil temperature and wetness. All of the forcing data are interpolated to the $1^\circ \times 1^\circ$ model resolution at the hourly interval.

We perform 10 sensitivity simulations with WFDEI reanalysis to distinguish driving factors for the changes in land carbon fluxes and BVOC emissions in the past 3 decades (Table 1). The control simulation (CO₂_MET_LUC) uses interannually-varying meteorology, [CO₂], and land cover for 1980–2011. The CO₂_MET run is the same as the control simulation but prescribes land cover at the year 1980. Three single-factor runs prescribe most boundary conditions at the year 1980 but allow the inter-annual variations of [CO₂] (CO₂_ONLY), land cover (LUC_ONLY), and meteorology (MET_ONLY) respectively. Results from these runs are compared with that of control simulation to determine the dominant drivers of simulated trends. To understand the impact of individual meteorological variables, three additional runs are performed with fixed (or recycled) [CO₂], land cover, and all meteorology at year 1980 but one field varying for 1980–2011 each time, including temperature (TEMP_ONLY), PAR (PAR_ONLY), and soil wetness (SOILW_ONLY). Finally, two runs are performed to examine feedback of biospheric changes. LAI_ONLY prescribes all boundary conditions at the starting year 1980 but implements the year-to-year LAI simulated by the control run. PHEN_ONLY also prescribes all forcings at the starting year except for the

estimated (by 87% in Amazon, 78% in North America, and 48% in Central Africa) or opposite in sign (East Asia and Indonesia) compared to observations. Such inconsistencies indicate the limit of model simulations, but may also in part result from the uncertainties in the satellite measurements (see Sect. 4.1).

3.2 Trends in land carbon fluxes

Predicted GPP and NPP trends show similar spatial pattern as that of LAI (Fig. 2a and b). However, regional trends are all positive in the main continents and on the global scale (Tables 2 and 3). Tropical areas are experiencing maximum changes, especially in Central Africa (GPP by 83.3 TgCa^{-2} and NPP by 51.7 TgCa^{-2}) and the Amazon (52.7 and 27.1 TgCa^{-2}). In the Northern Hemisphere (NH), changes are significant in Europe (53.4 and 33.2 TgCa^{-2}), East Asia (42.4 and 27.2 TgCa^{-2}), and North America (13.6 and 9.7 TgCa^{-2}). On the larger domain, the YIBs model predicts NPP trends of 67.4 TgCa^{-2} in northern land ($25\text{--}90^\circ \text{N}$) and 98.1 TgCa^{-2} in tropical land ($15^\circ \text{S}\text{--}25^\circ \text{N}$), similar to the ensemble estimates of 63 ± 22 and $102 \pm 34 \text{ TgCa}^{-2}$ for 1990–2009 based on 9 terrestrial biosphere models (Sitch et al., 2015). However, the simulated NPP trend is only 19.8 TgCa^{-2} in southern land ($15\text{--}90^\circ \text{S}$), much lower than the ensemble mean value of $53 \pm 31 \text{ TgCa}^{-2}$ in Sitch et al. (2015).

30 year historical observations of GPP and NPP are not available. Therefore, we compare YIBs predictions with MODIS land carbon fluxes over the more recent period of 2000–2011 (Fig. 3). Different from the 30 year trend, land carbon fluxes over the recent decade show negative trends in southeastern US, southern Africa, eastern Australia, and central and northern Asia (Fig. 3a and c). Most of these changes are consistent with the MODIS observations (except for the US, Fig. 3b and d) and are attributed to the drought tendency in the past decade (Zhao and Running, 2010). For the 30 year trend, both CO_2 and meteorology are playing important roles (Fig. 2d and e). CO_2 fertilization dominates the GPP and NPP trends of tropical forests in the Amazon, central Africa, and Indonesia, and ENF and DBF in boreal North America, eastern Europe, and central and northern Asia. Meteorological forcing drives changes in land

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carbon fluxes for tundra in subarctic regions, C3 grasslands in the central US, southern Africa, and the southern tips of South America, C4 grasslands in central Africa and the east of South America, and shrublands in Australia and southern Asia. Land use change plays a limited role in land carbon cycle flux trends over the past 3 decades, except for some areas in northern Africa.

We further distinguish sub-drivers for the meteorology-induced GPP trends (Fig. 4). Soil wetness plays the dominant role in the tropical and subtropical areas (Fig. 4c). The drought tendency in the western US, central Africa, and the east of South America (Fig. S2d in the Supplement) results in the regional decline of land carbon fluxes (Fig. 4a). In contrast, the increasing wetness in the northern Amazon and southern Africa leads to the enhancement of regional GPP. Warming is the main cause for the GPP trends over the subarctic areas (Fig. 4c). Contribution of PAR is limited, except for some areas in the eastern Europe. Driven with MERRA reanalysis, simulated GPP shows similar trends as that with WFDEI reanalysis, except for tropical areas where MERRA-forced runs exhibit stronger magnitude (Fig. 4b). The cooling tendency of MERRA temperature in the Amazon (Fig. S3a) increases regional GPP, because the climatological temperature is higher than the optimal temperature of 25 °C for the light-saturated photosynthesis (Yue and Unger, 2015). The large reductions in soil wetness in central Africa (Fig. S3d) result in the decline of GPP (Fig. 4b). In the western US, predictions with MERRA show smaller reductions in GPP relative to those using WFDEI, because the drought tendency is weaker in the former dataset (Figs. S2d and S3d).

The simulated net ecosystem productivity (NEP) shows weaker trends compared with GPP and NPP (Fig. 2c), because NEP is offset by the trends in heterotrophic respiration (Rh) (Table 2). Regionally, the YIBs model predicts enhanced net land carbon uptake in boreal North America, northern Asia, and southern Africa but reduced NEP in the central US, the Amazon, central Africa, eastern Europe, and East Asia. All these changes are consistent with the ensemble estimates based on 9 terrestrial models for 1990–2009 (Sitch et al., 2015). On the larger domain, the YIBs model predicts NEP trends of 2.0 Tg C a⁻² in northern land, 1.0 Tg C a⁻² in tropical land, and -0.3 Tg C a⁻² in

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southern land, much smaller in magnitude compared with the -2.0 ± 12 , 36.0 ± 13 , and $21 \pm 17 \text{ TgCa}^{-2}$ estimated by Sitch et al. (2015). However, their predictions are insignificant ($p > 0.05$) for 9, 5, and 7 out of 9 models in the northern, tropical, and southern land respectively, suggesting that the strengthening uptake by terrestrial ecosystem is not robust. Our simulated NEP trends (Fig. 5d) are in broad agreement with the comprehensive bottom-up estimates by Pan et al. (2011), who found slightly decreasing net carbon uptake by global established forests (without human perturbations in the tropics but with afforestation in subtropical areas) in 2000–2007 relative to that in 1990–1999. Attribution analysis shows that the NEP trends are mainly driven by the changes in meteorological forcings (Fig. 2f), because CO_2 fertilization enhances both NPP and Rh with similar magnitude (Fig. 5). A similar conclusion is achieved using MERRA reanalysis (Fig. S1f).

On the global scale, GPP, NPP, and Rh increase respectively by 298, 185, and 181 TgCa^{-2} in the past 3 decades (Table 3). The ensemble estimates by Sitch et al. (2015) yield a NPP trend of $218 \pm 76 \text{ TgCa}^{-2}$ and Rh trend of $160 \pm 53 \text{ TgCa}^{-2}$. The lower NPP trend in YIBs is mainly attributed to the smaller enhancement in southern land due to the negative trends in South America (Fig. 2b). The higher Rh trend in YIBs is a result of the larger enhancement in tropical forests due to the increased litterfall following CO_2 fertilization (not shown). The long-term trends of carbon fluxes are mainly driven by CO_2 fertilization, while the interannual variability is related to meteorological forcings (Fig. 5). Warming alone decreases GPP especially in tropical forests (not shown) but increases autotrophic respiration (Ra), leading to global reductions of 56 TgCa^{-2} in NPP and 10 TgCa^{-2} in NEP (Table 3). Drought alone strongly decreases GPP, especially for tropical grassland and shrubland (Fig. 4), leading to reductions of 51 TgCa^{-2} in NPP and 13 TgCa^{-2} in NEP. Trends in PAR do not affect GPP and NPP, but may decrease NEP by 23 TgCa^{-2} because soil respiration is slowly increasing to reach the equilibrium. Land use change has very limited impacts on the trends of carbon fluxes, though it induces relatively large reductions in NEP (Table 3). In this study,

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we do not include land-use-induced carbon emissions, which result in a net land carbon source (Ciais et al., 2013).

3.3 Trends in BVOC emissions

Simulated isoprene emission trends are sensitive to the choice of modeling scheme. With the PS_BVOC scheme, global isoprene emissions increase by 0.4 TgCa^{-2} during 1982–2011. Large enhancements are predicted in central Africa (0.25 TgCa^{-2}) and Europe (0.16 TgCa^{-2}), while moderate reductions are found in the western US, eastern South America, and East Asia (Fig. 6a). Drought accounts for the decline of isoprene emissions in the US and South America, but land use change is the main driver for the reductions in East Asia (Fig. 6d). Increasing $[\text{CO}_2]$ promotes photosynthesis but meanwhile inhibits BVOC emissions, leading to offsetting CO_2 effects on isoprene. Consequently, the global isoprene emission is mainly driven by meteorological changes (Fig. 6d). In contrast, using MEGAN scheme, the YIBs model simulates a global reduction of 1.1 TgCa^{-2} for isoprene emissions (Fig. 6b). Strong declines are found in the tropical rainforest, for example in the Amazon (-0.43 TgCa^{-2}), central Africa (-0.14 TgCa^{-2}), and Indonesia (-0.16 TgCa^{-2}) (Fig. 6b). The MEGAN scheme is sensitive to both light and temperature (Guenther et al., 1995). The strong positive brightening trends in PAR in Europe (Fig. S2b) promote isoprene emissions there. The positive impacts of NH warming (Fig. S2a) are compensated by CO_2 inhibition, leading to small changes in isoprene emissions (Fig. 6b). In the tropical areas, where trends of temperature and PAR are limited, CO_2 inhibition results in strong reductions of BVOC emissions. Monoterpene emissions show a global reduction of 0.04 TgCa^{-2} over the past 3 decades (Fig. 6c). Similar conclusions are achieved with the MERRA reanalysis (Fig. S4).

We compare our results with previous estimates of long-term BVOC emission changes. Stavrakou et al. (2014) investigated isoprene emissions over Asia during 1979–2012 using the MEGAN scheme and taking into account both climate and land-use changes. Their results showed widespread increases in the emissions over China

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but moderate decreases in Indonesia. In contrast, the YIBs model with the MEGAN scheme simulates widespread reductions in the same areas for 1980–2011 (Fig. 6b). The discrepancies between studies are accounted for by differences in the drivers including land cover change, meteorology, and CO₂ inhibition effects. The YIBs model is driven with land cover data from Hurtt et al. (2011), which estimates an increase of crop (non-isoprene emitter) fraction in East China by 0.32 % per year in the last 3 decades, at the cost of the coverage loss by 0.12 % a⁻¹ for DBF and 0.14 % a⁻¹ for ENF (strong BVOC emitters). However, the data from Ramankutty and Foley (1999), used by Stavrakou et al. (2014) with updates to 2007, show a reduction of the crop fraction over East China for the similar period. In addition, the ERA-Interim PAR used in Stavrakou et al. (2014) shows an increasing trend in southeast China (c.f. their Fig. 5c). On the contrary, the WFDEI PAR for YIBs exhibits a declining trend in the same region (Fig. S2b), leading to a reduction in isoprene emissions. The WFDEI surface solar radiation is based on the ERA-Interim radiation but is adjusted using the average cloud cover from the Climatic Research Unit (CRU) and taking into account the effects of interannual changes in atmospheric aerosols (Weedon et al., 2011). Finally, the YIBs simulations include CO₂ inhibition effects on BVOC emissions, which were neglected in Stavrakou et al. (2014).

Naik et al. (2004) predicted a global trend of 1.3 Tg C a⁻² for isoprene emissions during 1971–1990 using the Integrated Biospheric Simulator (IBIS) driven with monthly mean CRU meteorology. Lathiere et al. (2006) estimated an increasing global trend of 0.3 Tg C a⁻² for 1983–1995 using the ORCHIDEE (Organizing Carbon and Hydrology in Dynamic Ecosystems) vegetation model driven with sub-daily variables from the NCEP/DOE (National Center for Environmental Predictions/Department of Energy) Reanalysis 2. Muller et al. (2008) reported a global increase of 4.5 Tg C a⁻² for 1995–2006 using a canopy environmental model and the NCEP meteorological data. In contrast to these previous studies, YIBs with the MEGAN scheme simulates a decreasing trend of ~ 1 Tg C a⁻² in the past 3 decades. The main cause of the discrepancy in the sign of change is the missing CO₂ inhibition effects in the previous studies. In addition, differ-

ences in vegetation models, meteorological forcings, and time frames of investigation also likely contribute. The YIBs result is consistent with a recent study by Sindelarova et al. (2014), who reported a decreasing trend of $\sim 1.2 \text{ TgCa}^{-2}$ for global isoprene emissions during 1980–2010 using the MEGAN scheme and inclusion of a CO_2 inhibition parameterization from Heald et al. (2009).

3.4 Feedback of biospheric changes

Due to the changing climate and CO_2 fertilization, the biosphere is experiencing significant changes in the past 3 decades. The most evident alterations include LAI changes in peak season and phenological changes in growing and falling seasons. In this section, we explore the feedback of these biospheric changes to the carbon uptake and BVOC emissions.

3.4.1 Impacts of LAI changes

Sensitivity run LAI_ONLY retains the trends in LAI but prescribes other forcings. In this simulation, trends in GPP (Fig. S5a) and NPP (Fig. S5c) generally follow that in LAI (Fig. 1b), but with smaller magnitude relative to those in control simulations (Fig. 2a and b). LAI in the north of 30° N shows widespread increases in both observations and simulations (Fig. 1a and b). Over these northern lands, the unit change in leaf area leads to enhancement of regional GPP by 32 PgCa^{-1} , much lower than the response of $116 \text{ PgCa}^{-1} \text{ LAI}^{-1}$ for the simulation including CO_2 fertilization and climate forcings (Fig. 7a). In the tropical areas, both positive and negative LAI trends are predicted due to the competition between CO_2 fertilization and drought effects (Fig. 1). As a result, LAI-induced GPP and NPP changes show patchy distributions at tropics (Fig. S5a and S5c), leading to moderate changes in the global carbon assimilations (Table 3).

Trends in isoprene emission (calculated with the PS_BVOC scheme) also follow that of LAI, except that leaf expansion results in decreased emissions at high latitudes ($\sim 60^\circ \text{ N}$, Fig. S5e). The cause for such inconsistency is unclear, but might be because

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the denser leaves reduce radiation penetrating to lower canopy layers. Such impact would only affect BVOC emissions at high latitudes because PAR is usually limiting near subarctic areas. In most of the subtropical areas, increased LAI leads to enhanced isoprene emissions. On average, unit change in LAI at north of 30° N leads to enhanced isoprene emissions by 43 Tg C a⁻², only 25 % of the magnitude in simulation CO2_MET (Fig. 7b). A similar ratio of 23 % is achieved for MEGAN isoprene emissions. These results are consistent with that for GPP (Fig. 7a), suggesting that CO₂ fertilization and meteorological changes are the main drivers for the changes in carbon uptake and BVOC emissions, even over the northern lands where the most evident changes in LAI are observed.

3.4.2 Impacts of phenological changes

Plant phenology, which is the timing of budburst and leaf fall, is closely related to temperature, moisture, and photoperiod and thus is experiencing significant changes in the past decades following climate change (Jeong et al., 2011; Keenan et al., 2014; Buitener et al., 2015; Yue et al., 2015). Extension of the growing season has the potential to promote carbon uptake of forests (e.g., Piao et al., 2007; Richardson et al., 2009). Yet such inference requires careful interpretation because the phenological changes are usually accompanied with warming and elevated [CO₂], both of which are also contributing to the enhancement of carbon fluxes. Phenological changes are also expected to affect BVOC emissions, however, such investigations are still missing (Richardson et al., 2013). With the YIBs model, we evaluate the impacts of the growing season extension on both carbon uptake and BVOC emissions by isolating long-term phenological trends from changes in temperature and [CO₂].

The YIBs model simulates advanced spring and delayed autumn over most areas in NH (Fig. S6). Budburst dates advance on average by 0.16 days a⁻¹ in Europe and 0.15 days a⁻¹ in East Asia (Table 2), but with moderate changes or even delays in north-western Asia and eastern Siberia (Fig. S6a). Spring is earlier by 0.14 days a⁻¹ in eastern US while delayed by 0.15 days a⁻¹ in northwestern US and southeastern Canada,

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leading to a minor advance of 0.01 days a^{-1} over North America. Dormancy onset dates are largely delayed in eastern Europe and northwestern Asia ($\sim 0.3 \text{ days a}^{-1}$), western US ($\sim 0.1 \text{ days a}^{-1}$), boreal Canada ($\sim 0.1 \text{ days a}^{-1}$), and northeastern China ($\sim 0.1 \text{ days a}^{-1}$) (Fig. S6b). Advanced autumn ($\sim 0.1 \text{ days a}^{-1}$) is predicted in northern Asia. Most of these changes are consistent with observations from remote sensing data (Jeong et al., 2011), except for some discrepancies in the magnitude. The predicted phenological trends mainly follow the long-term changes of surface air temperature, especially that in April (for spring) and September (for autumn) (Fig. S7). Sensitivity tests without chilling requirement and photoperiod limit show similar changes (Yue et al., 2015), suggesting that temperature changes dominantly drive the trends of forest phenology in the past 3 decades.

On average, the YIBs model simulates advanced budburst by 0.12 days a^{-1} and delayed dormancy onset by 0.09 days a^{-1} at north of 30° N in the past 3 decades (Fig. 8a and b). Observations based on remote sensing greenness show trends of $-0.11 \text{ days a}^{-1}$ for onset and 0.25 days a^{-1} for offset during 1990–2009 (Zhu et al., 2013). An ensemble prediction based on 9 terrestrial models yields an advance of $0.08 \pm 0.13 \text{ days a}^{-1}$ for onset and a delay of $0.22 \pm 0.1 \text{ days a}^{-1}$ for offset (Sitch et al., 2015). Our predictions are in broad agreement with these estimates though the autumn delay is less, likely because the positive trend of offset is weaker for the recent decade (Jeong et al., 2011).

We plot the annual total GPP and isoprene emissions at north of 30° N against the length of growing season for 1982–2011 (Fig. 8c and d). In the $\text{CO}_2\text{-MET}$ run, the 1 day extension is correspondent to increases of 0.17 Pg C a^{-1} in GPP and 0.34 Tg C a^{-1} in isoprene emissions. If only temperature is allowed to vary, the phenological trend remains the same while the increases of GPP and isoprene emissions are largely weakened. In the TEMP-ONLY run, the 1 day extension in growing season is accompanied by increases of 0.05 Pg C a^{-1} in GPP and 0.25 Tg C a^{-1} in isoprene emissions. The changes in BVOC emissions are not as dramatic as those of GPP because CO_2 has both enhancing and suppressing impacts on the former. If we further exclude temper-

ature effects (PHEN_ONLY run), GPP increases only by 0.01 PgCa^{-1} while isoprene emissions decrease by 0.1 TgCa^{-1} , both of which are not statistically significant, suggesting that the phenological change alone does not promote either GPP or isoprene emissions. There are two reasons for this apparent contradiction. First, the extension of the growing season occurs in shoulder months, usually in May and September, when both GPP and BVOC emissions and their changes are much smaller compared to that in peak months (Fig. S8). Second, phenological changes are not uniform in space. As Fig. S6 shows, both positive and negative changes are predicted for budburst and dormancy onset dates. Such spatial inhomogeneity, in combination with the discrepancies in regional vegetation types and meteorological conditions, result in varied responses in GPP (Fig. S5b) and isoprene emissions (Fig. S5f).

Plant phenology at lower latitudes (30°S – 30°N) is also experiencing dramatic changes, though such changes are diverse in phase, magnitude, or both (Buitenwerf et al., 2015). In the model, tropical phenology is mainly driven by soil wetness and as a result exhibits large changes in the past 3 decades (not shown). These changes lead to a reduction of 42 TgCa^{-1} in GPP at the tropics (Fig. S5b), which accounts for 14% of global GPP trend but with the opposite sign (Table 3), suggesting additional inhibition of drought on carbon cycle. A similar conclusion applies for BVOC emissions (Fig. S5f), though experiments suggest that isoprene production has some tolerance to mild drought conditions (e.g., Pegoraro et al., 2006). However, changes in drought-dependent phenology are very uncertain and observations are not available for evaluation. We assume that phenological changes may have larger impacts on both carbon assimilation and BVOC emissions at tropical areas than that at higher latitudes.

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4.2 Impacts of CO₂ effects

Similar to the multi-model ensemble predictions (Sitch et al., 2015), we found that global trends in carbon fluxes are dominantly driven by CO₂ fertilization (Figs. 2 and 5). In the YIBs, the global responses to elevated [CO₂] is 0.2 %ppm⁻¹ for GPP and 0.27 %ppm⁻¹ for NPP, with relatively uniform spatial distribution (Fig. S10a and b). The GPP response falls within the range of 0.05–0.21 %ppm⁻¹ predicted by 10 terrestrial models (Piao et al., 2013) and that of 0.01–0.32 %ppm⁻¹ observed from multiple free-air CO₂ enrichment (FACE) sites (Ainsworth and Long, 2005). The NPP response is higher than the model ensemble of 0.16 %ppm⁻¹ (Piao et al., 2013) and the observed median value of 0.13 %ppm⁻¹ (Norby et al., 2005), suggesting that CO₂ fertilization to NPP may be overestimated in the YIBs. One possible cause is the omission of N limitation in the model, which could reduce CO₂ responses by half (Piao et al., 2013). Elevated [CO₂] leads to increases of 0.023 PgC a⁻¹ ppm⁻¹ in NEP, within the multi-model range of 0.003–0.06 PgC a⁻¹ ppm⁻¹ (Piao et al., 2013).

Responses of BVOC emissions to elevated [CO₂] are different between PS_BVOC and MEGAN schemes (Fig. S10c and d). PS_BVOC includes both CO₂ fertilization (on photosynthesis) and inhibition (on isoprene) effects, leading to moderate but generally positive changes in isoprene emissions. In contrast, emissions from the MEGAN scheme are not dependent on foliar photosynthesis and as a result only CO₂ inhibition is enforced. Chamber experiments show contrary tendencies for photosynthesis and isoprene in response to elevated [CO₂] (Possell et al., 2005), supporting the simulations with MEGAN. In addition, the magnitude of CO₂ inhibition implemented in MEGAN (–0.25 %ppm⁻¹) is close to observations (–0.26 %ppm⁻¹) in Possell et al. (2005). However, most of these experiments are conducted for short-term period and cannot detect LAI changes due to the long-term CO₂ fertilization. In addition, the impacts of CO₂ are dependent on species and environmental conditions (ambient temperature and light availability). For example, Buckley (2001) found almost no responses in isoprene emissions to the elevated [CO₂] for oak trees. Furthermore, experiments with

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iments, MEGAN scheme does not include drought inhibition on isoprene emissions. Simulations with YIBs show large responses of BVOC to soil wetness in tropical areas (Fig. S13d), mainly because of the changes in drought-dependent phenology.

Different climatic forcings may lead to varied trends in the simulations of carbon fluxes (Figs. 2 and S1) and BVOC emissions (Figs. 6 and S4). Trends of meteorology show large discrepancies between WFDEI and MERRA data sets over tropical areas, especially in Amazon and central Africa (Figs. S2 and S3). Such uncertainty strongly affects the calculated global trends in carbon budget and also influences the identification of drivers for those changes. For example, with MERRA forcing, YIBs model simulates global increases of 239 Tg C a^{-2} in NPP and 59 Tg C a^{-2} in NEP over the past 3 decades (Table S1 in the Supplement), both values are higher than that with WFDEI forcing (185 for NPP and 3 for NEP, Table 3) and much closer to the multi-model ensemble estimates (218 for NPP and 55 for NEP) as shown in Sitch et al. (2015). The stronger trends in carbon fluxes with MERRA are mainly attributed to the reductions in Amazonia temperature (Fig. S3a), which drives the large enhancement of regional GPP (Fig. 4b) and also overweighs the contribution of CO_2 fertilization (Fig. S1d). However, we are conservative about simulations with MERRA because the associated changes of meteorological fields are too strong in the tropics (Fig. S3). Our results show the large climate-driven uncertainties in the estimate of long-term trends in carbon fluxes, indicating the necessity of forcing inter-comparisons in addition to model inter-comparisons (e.g., Huntzinger et al., 2013; Piao et al., 2013).

4.4 Impacts of land use change

Changes of land use show moderate impacts on global carbon budget (Fig. 2) and BVOC emissions (Fig. 6) in the past 3 decades, though regional perturbations are found in China and Europe. The afforestation in Europe helps promote regional carbon uptake, resulting in more reasonable trends in LAI compared with remote sensing data (Fig. 1). However, the expansion of crop in China leads to a reduction in LAI, which is not supported by the satellite data. One possible cause is the uncertainty in crop frac-

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tion, because data from Hurtt et al. (2011), used by YIBs, show crop expansion while data from Ramankutty and Foley (1999) suggest reductions of the crop fraction over East China over the similar period. The role of land use change in our simulation might be conservative because we consider only land cover changes. Perturbed emissions from land use management, such as forest logging, cropping practice, use of fertilizer, fire management and so on (Houghton, 2010) may alter regional carbon budget by changing carbon sinks to sources. Studies including gross emissions of land use perturbation estimated a global net land source to atmosphere, which shows decreasing trend in the last 3 decades (Ciais et al., 2013). Such change may help strengthen net land carbon sink but is missing in our study.

4.5 Impacts of biospheric changes

The land biosphere has experienced significant changes in the past 3 decades. At north of 30° N, changes in LAI account for 25 % of the trends in regional carbon fluxes and isoprene emissions. However, the extension of growing season alone makes insignificant contributions to the increased carbon assimilation. This conclusion is inconsistent with site-level observations that show evident increases in carbon assimilation at early spring and/or late autumn in recent decades (Dragoni et al., 2011; Keenan et al., 2014). The causes for such discrepancies lie in two. First, phenology at specific location may exhibit much more intense changes than that at larger scale. For example, Dragoni et al. (2011) estimated extensions of growing season by 2.3–3.3 day a⁻¹ in Morgan-Monroe State Forest in south-central Indiana of US for 1998–2008. The magnitude of this change is ~ 10 times larger than the observed value of 0.36 day a⁻¹ from satellite and simulated value of 0.22 day a⁻¹ with YIBs for the northern lands. Second, enhanced temperature also contributes to the stronger uptake at early spring and late autumn. One difficulty for the observation-based estimate of phenological impacts is that extension of growing season is accompanied by warmer climate, which may stimulate both carbon assimilation and BVOC production. In a recent study, Barlow et al. (2015) found invariant length of land carbon uptake period at high northern

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latitudes based on the first time differential of atmospheric CO₂ concentrations, suggesting that increased greenness is not necessarily equal to enhanced carbon uptake in shoulder seasons. Furthermore, Barlow et al. (2015) showed that enhanced peak uptake is the main driver for the strengthened carbon sink at high northern latitudes over the past 4 decades. These conclusions are supportive of our simulations for the monthly trends at subtropical regions (North America, Europe, and East Asia) (Fig. S8).

5 Conclusions

With YIBs model, we estimated global increases of carbon assimilation especially at tropical areas for 1982–2011. This trend is mainly attributed to the widespread CO₂ fertilization effect, and jointly affected by changes in meteorology and land cover. Increase of temperature promotes carbon uptake of forest ecosystems at high latitudes (> 30° N) while drought tendency dampens GPP and NPP of grasslands and shrublands at low latitudes (30° S–30° N). The widespread increases of LAI at northern lands account for ~ 25 % of the regional GPP trends. Significant changes in phenology are found at north of 30° N; however, this temperature-driven phenological change alone is not promoting regional carbon assimilation. Changes in land use show limited influences on global carbon fluxes, except for some regional impacts over Europe (afforestation) and China (deforestation). Due to the simultaneous enhancement in soil respiration, land carbon sink has remained almost stable in the past 3 decades. In contrast, using alternate meteorological forcings, we found enhancement of land carbon sink due to increasing uptake in Amazon. The YIBs model does not yet include a fully coupled carbon-nitrogen cycle, thus the model may overestimate CO₂ fertilization effects. On the contrary, implementation of drought-dependent phenology may amplify drought inhibition effects on photosynthesis and result in an underestimation of carbon uptake.

We estimated global trends of BVOC emissions with two schemes. Simulations with PS_BVOC scheme show increasing isoprene emissions, mainly attributed to the increases of temperature. For this scheme, CO₂ effects are neutralized due to both fer-

tilization (on photosynthesis) and inhibition (on isoprene). Simulations with MEGAN scheme show decreasing emissions of isoprene and monoterpene because of CO₂ inhibition, especially in the tropics. In subtropical areas, both schemes predict regional increases of BVOC emissions in Europe following the warming trend and afforestation, but reductions in the US and China due to cropland expansion.

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Table 1. Summary of model simulations driven with WFDEI reanalysis.

Simulations	Descriptions
CO2_MET_LUC	Annually updated [CO ₂] and land cover, and hourly meteorology. All forcings vary for 1980–2011.
CO2_MET	Annually updated [CO ₂] and hourly meteorology for 1980–2011, land cover is prescribed at the year 1980.
CO2_ONLY	Annually updated [CO ₂] for 1980–2011, land cover is prescribed and hourly meteorology is recycled for the year 1980.
MET_ONLY	Hourly meteorology varies for 1980–2011. [CO ₂] and land cover are prescribed at the year 1980.
LUC_ONLY	Annually updated land cover for 1980–2011, [CO ₂] is prescribed and hourly meteorology is recycled for the year 1980.
TEMP_ONLY	Hourly temperature for 1980–2011 but other meteorological variables are recycled for 1980. [CO ₂] and land cover are prescribed at the year 1980.
PAR_ONLY	Hourly PAR for 1980–2011 but other meteorological variables are recycled for 1980. [CO ₂] and land cover are prescribed at the year 1980.
SOILW_ONLY	Hourly soil wetness for 1980–2011 but other meteorological variables are recycled for 1980. [CO ₂] and land cover are prescribed at the year 1980.
LAI_ONLY	Hourly meteorology is recycled for the year 1980. [CO ₂] and land cover are prescribed at the year 1980. Leaf area index varies for 1980–2011.
PHEN_ONLY	Hourly meteorology is recycled for the year 1980. [CO ₂] and land cover are prescribed at the year 1980. Phenology varies for 1980–2011.



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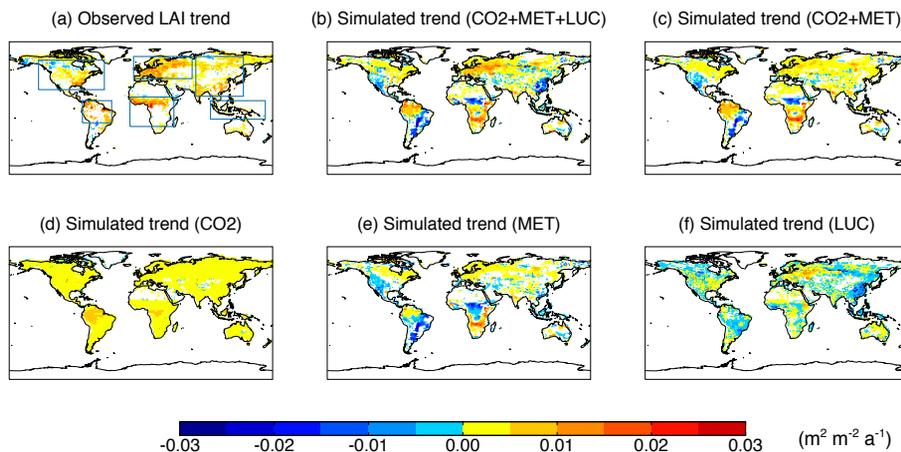


Figure 1. Comparison of trends in (b–f) simulated leaf area index (LAI) with (a) observations for 1982–2011. Observations are derived from GIMMS NDVI. Simulations are performed with either (d, e, f) single forcings or (b, c) the combinations of these forcings. Forcings considered include meteorology from WFDEI reanalysis (MET), CO₂ fertilization (CO₂), and land use change (LUC). For every forcing included in the simulation, the year-to-year fields are utilized. Otherwise, the forcing is prescribed at the year 1980. Only significant trends ($p < 0.05$) are presented. The six box regions in (a) indicate areas for statistical analyses in Table 2.

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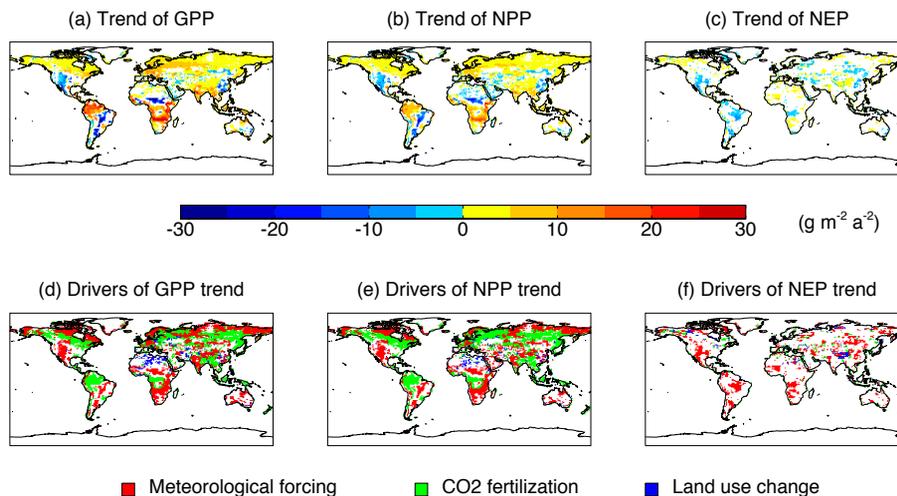


Figure 2. Simulated trends in **(a)** gross primary productivity (GPP), **(b)** net primary productivity (NPP), and **(c)** net ecosystem productivity (NEP), and **(d–f)** the dominant drivers for these changes during 1982–2011. Simulations are performed with WFDEI reanalysis. Three factors, meteorological forcing, CO₂ fertilization, and land use change, are considered as the potential drivers of flux trends. For each grid, the factor generating the largest (either maximum or minimum) trend with the same sign as the net change **(a–c)** is selected as the driving factor. Only significant trends ($p < 0.05$) are presented. Results for simulations with MERRA reanalysis are shown in Fig. S1.

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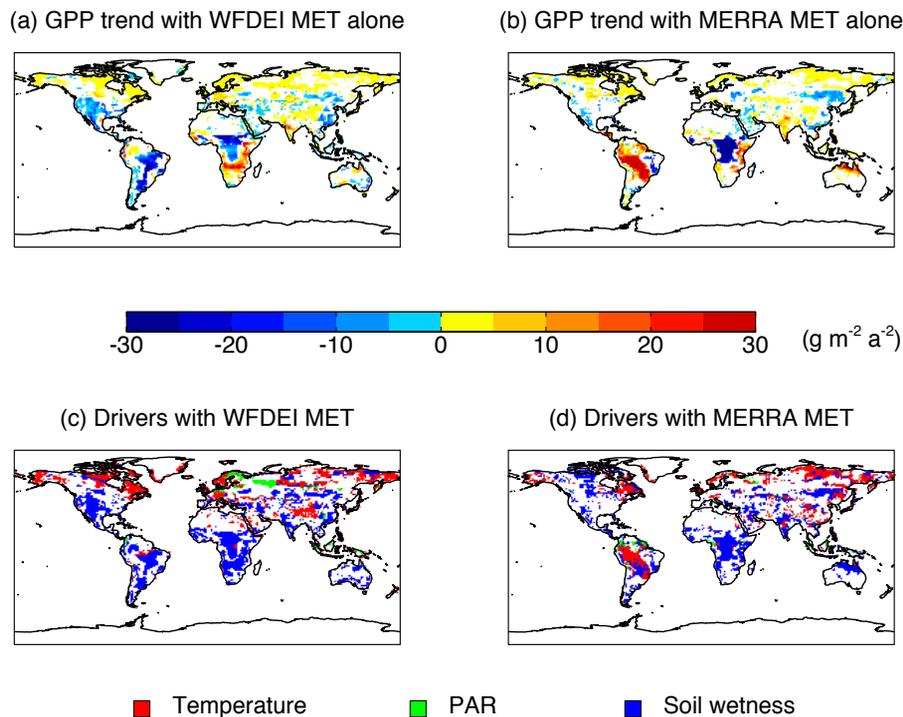


Figure 4. Simulated trends in GPP driven alone with **(a)** WFDEI and **(b)** MERRA reanalysis, and the **(c–d)** drivers for such changes. Simulations in **(a)** and **(b)** are performed with year-to-year meteorological forcings but prescribed $[\text{CO}_2]$ and land use in the year 1980. Simulations in **(c)** and **(d)** are the same as **(a)** and **(b)** respectively, except that the year-to-year variations are allowed only for a single meteorological variable (temperature, PAR, or soil wetness) each time. For each grid, the meteorological variable generating the largest (either maximum or minimum) trend with the same sign as the net change **(a–b)** is selected as the driving factor. Only significant trends ($p < 0.05$) are presented.

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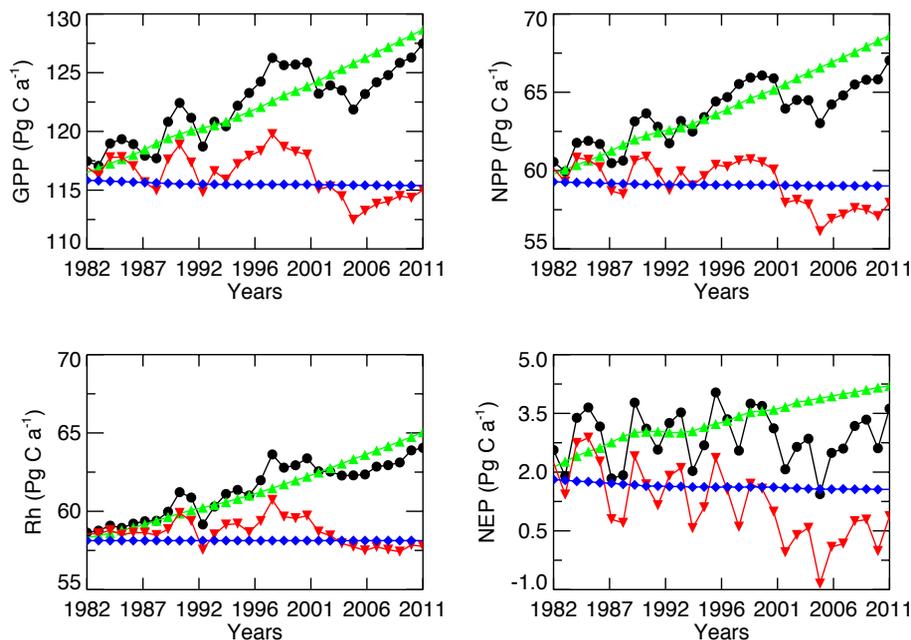


Figure 5. Global total fluxes of GPP, NPP, Rh (heterotrophic respiration), and NEP from different sensitivity simulations with all forcings (black), meteorology alone (red), CO₂ alone (green), and land use change alone (blue).

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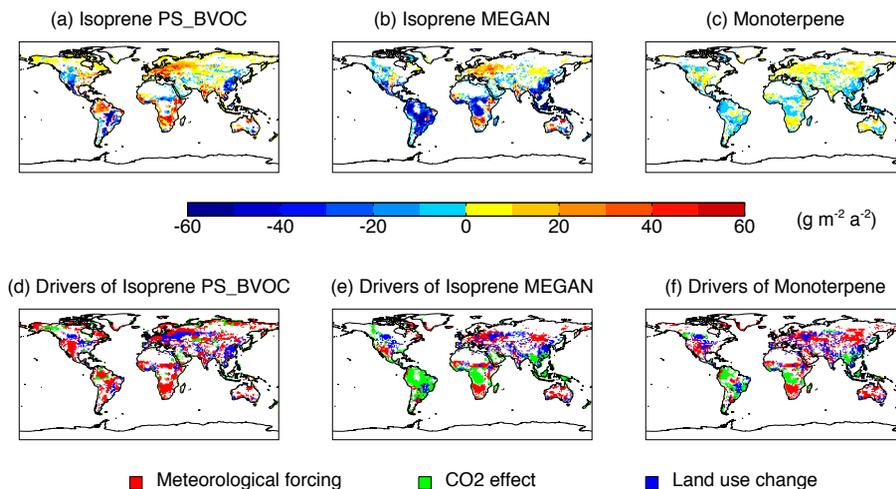


Figure 6. Simulated trends of (a, b) isoprene and (c) monoterpene, and (d–f) the dominant drivers for these changes during 1982–2011. Simulations are performed with WFDEI reanalysis. Isoprene emissions are simulated with (a) PS_BVOC and (b) MEGAN schemes. Three factors, meteorological forcing, CO₂ effects (both fertilization and inhibition), and land use change, are considered as the potential drivers of flux trends. For each grid, the factor generating the largest (either maximum or minimum) trend with the same sign as the net change (a–c) is selected as the driving factor. Only significant trends ($p < 0.05$) are presented. Results for simulations with MERRA reanalysis are shown in Fig. S4.

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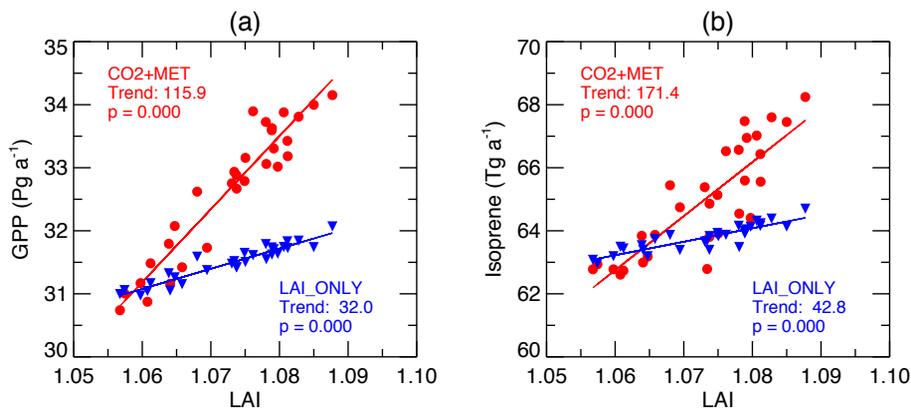


Figure 7. Responses of **(a)** GPP and **(b)** isoprene emissions to the changes in the annual average LAI at the north of 30° N for simulations CO2_MET (red) and LAI_ONLY (blue). Both GPP and isoprene emissions are the sum of all PFTs. Isoprene is simulated with the PS_BVOC scheme. Units of trends are **(a)** PgC a⁻¹ LAI⁻¹ and **(b)** TgC a⁻¹ LAI⁻¹. The spatial distribution of GPP and isoprene changes is shown in Fig. S5.

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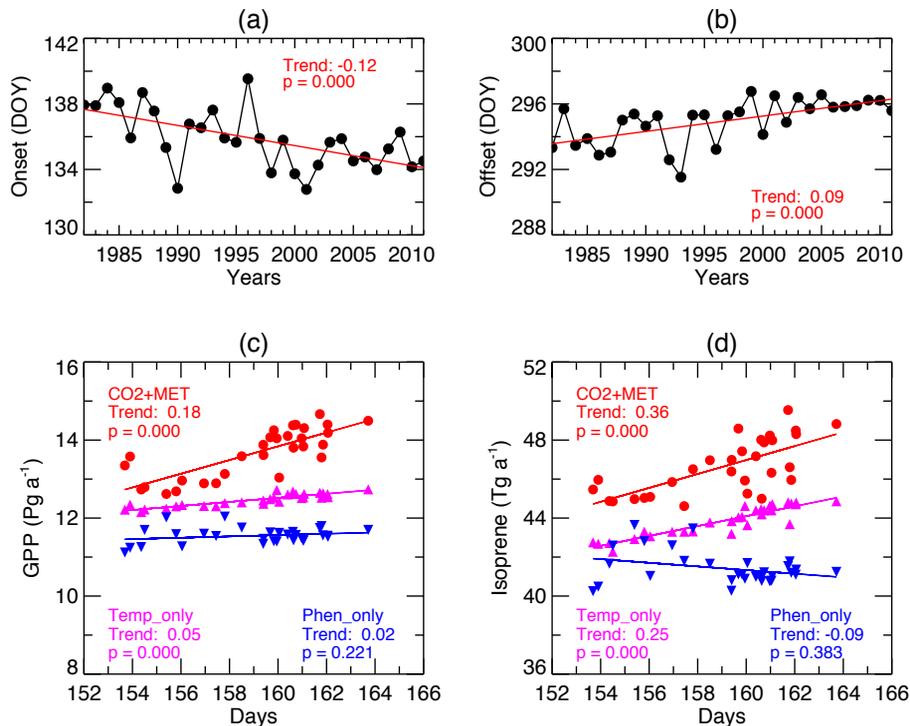


Figure 8. Predicted trend in **(a)** budburst and **(b)** dormancy onset dates over north of 30° N and the responses of **(c)** GPP and **(d)** isoprene emissions to the changes in the growing length. Both GPP and isoprene emissions are the sum of DBF, shrub, grassland, and tundra. Isoprene is simulated with the PS_BVOC scheme. For the bottom panel, different colors indicate sensitivity experiments with different year-to-year forcings: CO_2 and meteorology (red), temperature only (magenta), and phenology only (blue). Units of trends are **(a)** day a^{-1} , **(b)** day a^{-1} , **(c)** $\text{Pg C a}^{-1} \text{ day}^{-1}$, and **(d)** $\text{Tg C a}^{-1} \text{ day}^{-1}$. The spatial distribution of GPP and isoprene changes is shown in Fig. S5.

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