Terrestrial biogeochemical feedbacks in the climate system

Supplementary material

Figure 1

Using the carbon cycle CO₂- and climate response, Gregory et al.¹ presented a formalised analysis that can serve as the basis to compare results across different model studies that investigate different forcing agents and apply different scenarios, by using normalised feedback factors (W m⁻² K⁻¹) analogous to feedback calculations in studies of the physical climate system. In Figure 1, we propose an expansion of the Gregory et al. analysis to a number of additional processes discussed in this review. The analysis requires linearization of the problem¹ which introduces uncertainties in the presented numbers. Given the non-linearity of many biogenic and atmospheric processes, particularly those related to shorter-lived compounds, and noting that coupled model simulations do not yet exist for most of the discussed processes it is thus imperative to note that Figure 1 can only be interpreted as a first approximation of the relative magnitude of biological processes in the climate system. Moreover, the values shown are not constants and only apply to gradual changes in climate (using here a change in mean temperature as surrogate) over the 21st century. The total radiative forcing drawn in Figure 1 assumes additive behaviour of individual forcings, the maximum representing the top-end of all estimates including (i) the maximum from C-only models, and (ii) results from the C-N simulations (black open rectangle). This ignores additional possible non-linear (but unknown) interactions that may be either increasing or decreasing the total forcing. Some important processes (e.g., feedbacks from terrestrial N₂O emissions or biogenic SOA) could not be included, even as an initial estimate, due to lack of suitable studies to date.
1) Effects of enhanced land carbon uptake by CO₂ fertilisation (range -0.17 to -1.9 Wm⁻²K⁻¹; in white: average and standard deviation of the C4MIP² models: -1.1 ± 0.5 Wm⁻²K⁻¹; vs. the terrestrial carbon cycle-climate feedback² (range 0.1 to 0.9 Wm⁻²K⁻¹; 0.4 ± 0.2 Wm⁻²K⁻¹). In the C4MIP intercomparison², all models calculated a higher atmospheric CO₂ concentration in coupled compared to uncoupled simulations, caused by a reduction of terrestrial and marine carbon uptake as a result of climate change; changes on land dominated the response in most models. Based on Gregory et al.¹ (their equations 6 and 20, but concentrating on the land component) the radiative forcings (RF) due to CO₂ fertilisation vs. those due to carbon loss due to increased respiration and the temperature response of carbon assimilation can be expressed as

\[ r_{\beta, CO₂} = \rho \beta_{land, CO₂} \] (1)

\[ r_{\gamma, CO₂} = \phi \gamma_{land, CO₂} \] (2)

with the values of \( \beta_{land, CO₂} [\text{ppmv ppmv}^{-1}] \) and \( \gamma_{land, CO₂} [\text{Pg C K}^{-1}] \) taken from Table 3 in ². \( \rho \ [\text{W m}^{-2} \text{ K}^{-1}] \) is the climate resistance of a specific C4MIP model, taken as the radiative forcing from CO₂ over the change in global mean temperature in 2100 ¹. \( \phi \) represents the linear approximation of the increase in RF with increasing atmospheric concentrations (0.01 W m⁻² ppmv⁻¹ and 0.0049 W m⁻² PgC⁻¹, respectively. For small changes in atmospheric CO₂ this simplification can be used to substitute the non-linear relationship¹:

\[ F = F_{2x} \ln \left( \frac{C(t)}{C(0)} \right) / \ln(2) \approx \phi (C(t) - C(0)) \] (3),

where \( F_{2x} \) is RF resulting from doubling of atmospheric CO₂, and \( C(t) \) and \( C(0) \) are atmospheric CO₂ burdens at the end and beginning of a scenario (or coupled model simulation), respectively³.

2) Recently, radiative forcings from carbon-cycle climate models explicitly accounting for terrestrial N dynamics have been derived in a number of studies⁴-⁶. Numbers in Figure 1 were calculated as in (1) using the reported values for \( \beta_{land, CO₂, N} \) and \( \gamma_{land, CO₂, N} \) in 2100. Due to N
limitation, simulations with coupled carbon-nitrogen cycles show a strong reduction of the CO₂ fertilisation effect (-0.4 to -0.8 Wm⁻²K⁻¹) in comparison to models that do not account for terrestrial nitrogen dynamics. This reduced the terrestrial C accumulation in response to rising atmospheric CO₂ concentration and yields a higher atmospheric CO₂ burden at the end of the 21st century. Increased soil N mineralization stimulates plant growth in all models, however, magnitude and geographic location of this effect varies between a globally minor effect and a transient compensation for temperature related global land C losses (-0.1 to 0.25 Wm⁻²K⁻¹). Nevertheless, for the two model systems that have reported both C and C-N simulations⁴,⁶, accounting for the synergistic interactions of changing climate, CO₂ concentration and N deposition consistently results in higher atmospheric CO₂ levels in C-N compared to C-only simulations with the same model framework. In these calculations, RF range from 0.23 to 0.47 W m⁻²⁴,⁶ as the N limitation effect on CO₂ fertilisation is the strongest effect.

3) Radiative forcing from 21st century carbon loss in permafrost melt and global wetlands as a result of global warming. A maximum value was derived from adopting Gruber et al's⁷ estimates of a possible release of 200 Pg C at the end of the 21st century, which is unlikely but not completely implausible compared with some recent regional estimates⁸-⁹. The minimum estimated value in the Figure was taken from a coupled terrestrial ecosystem-climate model experiment restricted to northern ecosystems¹⁰ who found between 7 and 17 Pg C emitted from wetlands and permafrost for a range of climate and CO₂ scenarios. For simplicity, we assume here that the entire C released is in the form of CO₂, and that 50% of this additional carbon is being absorbed by the ocean (following¹¹, see (5)). Radiative forcing calculations (0.001 to 0.24 Wm⁻²K⁻¹) were done according to eq. (2), assuming that the released CO₂ is additional to the land C loss by the year 2100 simulated in the coupled simulations in², and taking the mean change in temperature from
these studies as $dT$. Synergistic effects of the added CO$_2$ on the CO$_2$ fertilisation effect were thus not taken into account.

4) Radiative forcing in 2100 relative to present-day or past base values from enhanced methane emissions from wetlands in response to anthropogenic warming$^{12-15}$. We adopt the ideas of Gregory et al. $^1$ (see 1) and search for the relationship between the radiative forcing of methane, and the change in atmospheric burden of methane following global warming related changes in methane emissions in a linearized form (following Gregory et al’s argument that for small changes in atmospheric burden such a linearization would be at the cost of relatively small errors).

Equivalent to eq. (3), the RF for methane is approximated as$^3$

$$RF_{CH4} = 0.036*(CH_4^{0.5} - CH_4^{BASE}^{0.5}) – f(CH_4, N_2O)$$ and $$RF_{CH4} \approx \phi \frac{d[CH_4]}{dT}$$ (4),

Where $f(CH_4, N_2O)$ is a typically applied correction term, given e.g., in $^3$ and $\phi = 0.036*[(CH_4,2100^{0.5} - CH_4^{BASE}^{0.5}) - f(CH_4, N_2O)]/(CH_4,2100-CH_4^{BASE})$. Values for future CH$_4$$_{2100}$ and present or past CH$_4$$_{BASE}$ were taken from$^{12-14}$.

To calculate the change in atmospheric burden of methane due to a unit increase in global mean temperature, we first approximate the cumulative emissions of methane from wetlands by the change in global mean temperature, ie.

$$\gamma_{methane} = \int (e^{CH_4(t)} - e^{CH_4(1990)}) \frac{dt}{dT}$$ (5).

This approximation seems appropriate for all studies used in Figure 1, as the global emissions in the simulations increase rather steadily with time and temperature, and since those studies that use a pre-industrial starting point have shown only small changes in emissions to take place before 1990. To account for interactions of methane emissions with atmospheric chemistry as simulated by the different models, we then approximate the change in atmospheric burden by relating the cumulative emissions to the simulated change in burden.
\[ \theta_{\text{methane}} = \frac{dCH_4_{\text{atm}}}{\int (e^{CH_4(t)} - e^{CH_4(1990)}) \, dt} \quad (6), \]

which leads to rather consistent results across the variety of scenarios and models considered. The radiative forcing (0.02 to 0.1 W m\(^{-2}\) K\(^{-1}\)) is then approximated as \( RF = \phi_{\text{methane}} \theta_{\text{methane}} \gamma_{\text{methane}} \, dT \), leading to the following equation for the methane feedback parameter:

\[ r_{\text{methane}} = \frac{RF}{dT} = \phi_{\text{methane}} \theta_{\text{methane}} \gamma_{\text{methane}} \quad (7). \]

5) Radiative forcing estimates from the additional atmospheric CO\(_2\) burden arising from ozone toxicity effects on plant photosynthesis and the ensuing reduced land carbon uptake\(^1\). Calculations were based on the IPCC A2 high anthropogenic precursor emission scenario. Sitch et al\(^1\) estimated for the year 2100 (compared to baseline levels of 299 ppm in 1901, the start of the simulation) a maximum indirect forcing of 0.62 and 1.09 W m\(^{-2}\) for low and high sensitivity of plants to ozone, with a concomitant forcing calculation that seeks to treat CO\(_2\) and O\(_3\) effects jointly (\(^1\), their Supplementary information). The additional atmospheric CO\(_2\) burden was derived from differences in land carbon storage between experiment \( \Delta CO_2 \) and fixed O\(_3\) vs. \( \Delta CO_2 \) and \( \Delta O_3 \) (see Table 1, Sitch et al.), and assuming 50% of the extra CO\(_2\) from land carbon changes to be sequestered into oceans (analogue to Sitch et al.). To convert these results into a climate feedback parameter and the units used in Figure 1 (0.08 to 0.17 W m\(^{-2}\) K\(^{-1}\)), we first calculate the different cumulative changes in land carbon storage for these simulations relative to the change in atmospheric CO\(_2\), to arrive at a modified value for \( \beta_{\text{land}, CO_2} \) taking account of the Ozone effect (\( \beta_{\text{land}, CO_2, O_3} \)). We then apply eq (1) to quantify the feedback parameter associated with the phytotoxic effect of ozone (\( r_{\beta, CO_2, O_3} \)), using for consistency reasons the value of \( \rho \) from HadCM3 in \(^2\).

6) Effects of varying BVOC emissions on estimated global total ozone radiative forcing. Limited late 19\(^{th}\) and early 20\(^{th}\) century atmospheric observations support the idea that past O\(_3\)
burden was lower than is generally inferred from simulation studies, although methodological
deficiencies suggest that these observations must be treated with caution\textsuperscript{16-17}. Based on studies that
assumed variable pre-industrial (PI) BVOC and NO\textsubscript{x} emissions\textsuperscript{16-17} we calculated an enhanced PI
to present RF by 0.03 to 0.12 W m\textsuperscript{-2} (not shown), compared to a standard simulation with pre-
industrial and present biogenic emissions assumed to be unchanged. The uncertainty in present day
O\textsubscript{3} forcing due to presence or absence of isoprene adds another several W m\textsuperscript{-2} (ca. 0.09\textsuperscript{18}).

In Figure 1 we show the additional forcing arising solely from changed present to future BVOC
emissions in response to warming\textsuperscript{19} which therefore is likely lower compared to a pre-industrial to
future analysis. Ignoring the possible additional effects due to a direct CO\textsubscript{2} interaction\textsuperscript{20}, various
studies\textsuperscript{21} have suggested that global temperature effects would increase BVOC emissions from
terrestrial ecosystems from present-day values by about 7 to 15\% per degree K warming, although
modelling studies that combine effects of climate change, and associated changes in vegetation
productivity suggested that this increase could also be in the order of 50\% K\textsuperscript{-1}\textsuperscript{19,21-22}. Using an
estimate of global present-day BVOC emissions of 696 Tg C a\textsuperscript{-1}\textsuperscript{22}, suggests a climate sensitivity
\(\gamma_{BVOC}\) of 48 to 348 Tg C\textsubscript{BVOC} a\textsuperscript{-1} K\textsuperscript{-1}. Since the atmospheric lifetime of both BVOC and O\textsubscript{3} is short,
we relate emissions directly to changes in ozone burden, i.e.

\[
d[O_3] = \theta_{ozone} \, dBVOC \quad (8).\]

\(\theta_{ozone}\) depends on magnitude and spatial pattern of emissions and the rate of ozone destruction
in the atmosphere. As a first-order approximation, results from the “AB” and “ABC” experiments
in\textsuperscript{19} allow to quantify this as 0.023-0.035 Tg [O\textsubscript{3}] Tg C\textsubscript{BVOC} a\textsuperscript{-1}. Note that this estimate is based on
simultaneously changing biogenic NO\textsubscript{x} emissions. Hauglustaine et al.\textsuperscript{19} further calculate the
radiative forcing based on the simulated changes in O\textsubscript{3} burden from which we derive our
approximation as \(\phi_{ozone} = 0.004-0.005\) W m\textsuperscript{-2} Tg\textsuperscript{-1} O\textsubscript{3}. However, this estimate should be taken with
due caution, as the radiative effect may vary strongly geographically, and does not necessarily add
up linearly to global forcing totals. Assuming that the spatial patterns of changes in BVOC emissions are comparable between simulations, we apply this value also minimum to maximum range of the studies in Carlslaw et al.\textsuperscript{21} and arrive at an estimate of the effect of BVOC on the radiative forcing of ozone as

$$r_{ozone} = \frac{R_{O3,BVOC}}{dT} = \phi_{ozone} \theta_{ozone} \gamma_{BVOC} \quad (9).$$

These rough estimates (0.004 to 0.06 Wm\textsuperscript{-2}K\textsuperscript{-1}) disregard the additional effects of warming on biogenic emissions of NO\textsubscript{x} emissions, but too few studies to this effect are available to date to quantify the likely effect on global radiative forcing.

7) Fire: Pyrogenic emissions can have a cooling (e.g., SO\textsubscript{4} or organic carbon aerosol) or warming (e.g., black carbon, O\textsubscript{3} and CH\textsubscript{4}) effect in the atmosphere\textsuperscript{23}. Present biomass burning is mostly associated with human activities in tropical regions, often related to land clearing. Estimates of land use change emissions are incorporated into SRES emission scenarios and thus do no fully fall into our adopted framework of biogenic processes and feedbacks in the climate system. In addition to net carbon loss from land cover change driven by fire, Bowman et al\textsuperscript{24} have estimated small net radiative forcing from fire emissions from the pre-industrial to present as the result of both cooling and warming substances emitted from biomass burning. A number of regional studies indicate a potentially strong response of area burnt and fire emissions due to climate change, directly via burn conditions and lighting ignition sources, and indirectly via vegetation productivity and plant litter available for combustion\textsuperscript{24-25}. Studies summarised in Flannigan et al.\textsuperscript{25} indicate that climate change effects alone may increase area burned up to 3 times (2XCO\textsubscript{2} scenarios) or even 5 times and more (3XCO\textsubscript{2} scenarios) compared to present. When neglecting possible additional effects of a changing fire intensity or fuel load, changes in area burnt can be used to scale future emissions\textsuperscript{21}. A factor of 2.4 was applied to link the change in pyrogenic aerosol emissions (A\textsubscript{pyr}) to a change in aerosol concentration [A\textsubscript{pyr}]\textsuperscript{21}. It is unknown
whether the linearization applied in (1) also applies for fire emissions. If we assume it to hold, we can propose analogue to the above (eq. 9):

\[ r_{A_\text{pyr}} = \frac{RF_{A_\text{pyr}}}{dT} = \phi_{A_\text{pyr}} \theta_{A_\text{pyr}} \gamma_{A_\text{pyr}} \]  

(10),

with

\[ \phi_{A_\text{pyr}} = \frac{RF_{A_\text{pyr}}}{d[A_\text{pyr}]}, \quad \theta_{A_\text{pyr}} = \frac{d[A_\text{pyr}]}{d(A_\text{pyr})}, \quad \text{and} \quad \gamma_{A_\text{pyr}} = \frac{d(A_\text{pyr})}{dT}. \]

Future RF can be obtained by scaling the estimated present day range (-0.05 to 0.22 W m\(^{-1}\); \(^{21}\)) to future concentrations; numbers in Figure 2 are from \(^{21}\) with a maximum of 300% imposed to increases in burnt area, and dT of 2 to 4.5K \(^{26}\). The derived estimates (-0.03 to 0.06 Wm\(^{-2}\)K\(^{-1}\)) do not include the additional warming effects of pyrogenic trace gas emissions\(^{23}\).

**Effects of RF by SOA in response to variable BVOC emissions**

A limited number of studies to date have attempted to isolate effects of varying global BVOC emissions not only on global SOA burden but also on the associated radiative forcing. In these experiments, interactions were investigated mostly from a model sensitivity perspective, and do not include future scenario calculations which prevent analysis based on the framework of \(^{1}\).

Present day radiative effects due to SOA were estimated with the global aerosol climate model ECHAM5/HAM (\(^{27}\), with extensions for prognostic cloud droplet number concentrations and ice crystal number concentrations by \(^{28}\)). The SOA scheme is based on that of Tsigaridis and Kanakidou\(^{29}\), extended to deal with the size-resolved aerosol microphysics in ECHAM5/HAM. The scheme includes isoprene and monoterpenes as well as anthropogenic aromatics as SOA precursors. Emissions of isoprene were calculated using the MEGAN scheme\(^{30}\) and of monoterpenes following Guenther et al (1995)\(^{31}\), with annual total emissions of 446 TgC isoprene and 89 TgC monoterpenes. Simulations of present day climate were made, nudged to ECMWF
reanalysis for the year 2000, with and without SOA. The shortwave radiative effect of SOA, calculated as the difference of the radiative fluxes at the top of the atmosphere, was \(-0.31\) Wm\(^{-2}\) for clear sky and \(-0.08\) Wm\(^{-2}\) all sky\(^{32}\). The indirect effect of SOA is thus positive in this model in the global annual mean. Regionally, this positive influence is seen mainly on persistent marine cloud decks that are affected by anthropogenic pollution, whether from industry or from biomass burning, whereas the indirect effect over pristine forest and ocean is generally weak. This behaviour arises from the model assumptions that (i) organics do not directly influence the formation of new particles throughout the troposphere, and that (ii) based on a simple extension of the partitioning theory of Pankow\(^{33}\) to size-resolved modes, gas-particle equilibrium of semi-volatile SOA favours the larger modes. The former means that in this model, SOA does not add to the particle number concentration, and the latter means that SOA condenses preferentially on particles that are already large enough to become CCN, while strengthening the coagulation sink of small particles. Where enough large particles are present due to pollution, this enhanced sink is strong enough to decrease the number of smaller particles, resulting in a reduction in the number of cloud droplets and a positive indirect effect.

Clearly, a discrepancy exists between process-descriptions made in today’s SOA models and results from recent studies and field observations that indicate CCN production due to condensation of organic vapours on aerosol particles to yield significant indicating negative indirect forcing \(^{34-35}\).

A second set of simulations was done with a similar SOA modelling scheme\(^{29}\), implemented into the GISS ModelE\(^{36}\). SOA sources include isoprene and monoterpenes oxidation products\(^{29}\) with the BVOC emission sources as in\(^{37}\). ORVOC sources (which include sesquiterpenes and higher terpenes) are lumped on monoprene production, and scaled from values in\(^{22}\) (Tsagaridis and Menon, unpublished). Anthropogenic SOA from aromatics is under development and thus
SOA mass and forcings estimated here are expected to be underestimated by about 10%. Applying a lower BVOC emission source than in the ECHAM/HAM study, the direct forcing from SOA aerosols was \(-0.02\) Wm\(^{-2}\) (obtained from differences between simulations with and without SOA aerosols) and the total aerosol direct forcing was \(-0.21\) Wm\(^{-2}\) from pre-industrial to present-day. We also included the interactions of total SOA aerosols with clouds through the aerosol-cloud interaction scheme in ModelE, as described in Menon et al.\(^{38}\) and further improved in order to take into account SOA aerosols. When simulating both the first and second aerosol indirect effects we estimate a total aerosol indirect effect of \(-1.17\) Wm\(^{-2}\); separate estimates of the BVOC-SOA contribution to the total aerosol effect are not available.

**Figure 2:**

Superposed epoch analysis (SEA\(^{39}\)) is an approach that reveals the consistent response of a time series to the occurrence of events of a particular nature (key events), like the episodes of abrupt warming that define the Dansgaard-Oeschger (D-O) “cycles”\(^{40-41}\). The analysis involves using the key events to align segments of an appropriately detrended \(^{42}\) time series, and then summarizing the aligned segments in a set of bins or averaging windows arranged around the events. If there is a consistent response, then this emerges as a distinct pattern; if not, the individual events cancel one another out, and no pattern appears. We examined the response of (a) the NGRIP (Northern Greenland Ice Core Project\(^{43}\)) Greenland ice core d\(^{18}\)O record (as an index of temperature), composite ice-core records of (b) CH\(_4\), (c) CO\(_2\), (d) N\(_2\)O, and (e) dust concentration (an index of vegetation cover, with more vegetation meaning less dust erosion and transport) in the NGRIP Greenland ice-core record, and (f) a composite curve of transformed and standardized charcoal concentrations (a proxy for biomass burning\(^{46}\)) from a global network of terrestrial and marine
cores. The ice-core data (Byrd: Byrd station; EDC: EPICA Dome C\textsuperscript{44,45}) were provided by Eric Wolff\textsuperscript{47}; the charcoal data is from Daniau et al.\textsuperscript{41}.

We detrended the individual time series using a locally weighted regression curve with a window half-width of 5000 years (to remove the impact of orbital time-scale variations), and then summarized the data by taking the median of data values that fell into 200-yr wide non-overlapping bins, a bin-width value appropriate for the lower-resolution series like the ice-core carbon dioxide record. We also analyzed the higher-resolution records using smaller bins as appropriate. The significance of the emergent SEA patterns can be assessed using a Monte Carlo approach in which the analysis is repeatedly performed using randomly chosen key-event times. Here we used 2000 repetitions to generate a 95-percent confidence interval for the null hypotheses of no characteristic response for each curve\textsuperscript{48}. The curves for the common 200-year bin-width analyses are plotted in colour on Figure 2, and the upper and lower confidence intervals are plotted as thin quasi-horizontal black lines. Curves for analyses using smaller bin widths for the higher resolution series are plotted as thin black lines. Shading indicates significant patterns in the response of the time series to the abrupt-warming and rapid-cooling events. The CO\textsubscript{2} and N\textsubscript{2}O records are much shorter, and of lower resolution, than the other ice-core records, and as a result it is more difficult for the SEA composite to achieve significance.

The Greenland oxygen-isotope record (Figure 2a), an index of regional temperature, shows the characteristic saw-tooth pattern of an individual D-O cycle, and all of the other records show distinctive responses associated with the occurrence of abrupt warming or cooling. The responses of all time series to the events are non-linear (linear responses would appear as inverted mirror images). The individual responses to abrupt warming can be categorized as abrupt and rapid with no appreciable lag, such as those for methane (b), dust (c), and biomass burning (f), all influenced by the hydrological status of the land surface, the nature of the vegetation cover, and vegetation
productivity, and progressive such as those for CO$_2$ (c) and N$_2$O (d), with more interconnected biogeochemical cycles. The responses to rapid cooling are similarly mixed, with those for CH$_4$ and dust again abrupt and rapid with no lag, and those for CO$_2$ and N$_2$O more gradual. There is an initial response (decrease) of biomass burning to cooling, followed by a gradual recovery.

As in the example of peatland growth and CH$_4$ discussed in the main text, the changes in these records illustrate the potential for amplification of the climate changes represented here by the coupled atmospheric-oceanic circulation reorganizations of the D-O cycles. For example, the abrupt increases in dust associated with D-O cooling events and their impact on dust-retaining vegetation cover, lead to further or prolonged cooling through the direct negative radiative forcing by increases in optical depth and land-surface albedo, and through the indirect effects of fertilization by dust of terrestrial and marine ecosystems which favour carbon sequestration. Increases in biomass burning in response to abrupt warming would likewise favour further warming by increased emissions of long-lived greenhouse gases, ozone and its precursors and black carbon. Although there may be a small negative radiative-forcing contribution by fire through increases in land-surface albedo, post-fire revegetation is generally rapid, and this contribution is likely very small relative to that of the emissions. Other work has shown$^{29,33}$ that the D-O cycles are registered throughout the climate system, and SEAs of this collection of records shows that the feedback to climate from the terrestrial and marine biosphere changes is generally positive, amplifying the initial climate changes.

Panel description, Box 1:

In each panel, atmospheric composition is indicated by ovals, climate (i.e. temperature) responses by diamonds. Biogeochemical processes responding to atmospheric drivers that can in turn feed back on atmospheric composition are shown as squares. It is this change in atmospheric
composition what ultimately feeds back to climate; we do not draw feedbacks on climate explicitly in most of the panels to keep these as simple as possible. We also omit effects other than temperature, although in many cases feedbacks via precipitation and radiation, and additional indirect effects via changes in vegetation composition and leaf area index also play a major role.

(I) Climate warming in response to anthropogenic emissions alters biogeochemical cycling (BGC) of a compound that directly acts as radiative forcing agent (e.g., a well-mixed greenhouse gas).

Panel a: Terrestrial carbon cycle-climate feedback. Increasing atmospheric CO$_2$ concentration warms climate. Increased temperatures stimulate ecosystem respiration (R) which further enhances CO$_2$ concentration (and temperature.) Photosynthesis (PS) is also a temperature-sensitive process, with a temperature maximum that varies between growth environments. Warmer temperatures can thus either enhance or reduce photosynthesis, depending on location and how close a plant operates to this maximum. Furthermore, both respiration and photosynthesis can also acclimate to changes in temperature. Feedbacks between the carbon cycle and the climate system are referred to as carbon-climate feedback, and the available carbon-cycle climate models suggest consistently this feedback to be positive. Enhanced CO$_2$ can also stimulate carbon assimilation; CO$_2$ uptake by higher net primary productivity (NPP) reduces atmospheric CO$_2$ growth rate, and thus the rate of climatic changes due to fossil fuel emissions. This is generally referred to as carbon-concentration feedback. The net between stimulated NPP and heterotrophic respiration determines whether the net land carbon sink declines$^2$.

Panel b: Warmer temperatures could stimulate soil mineralization making nitrogen available to plants$^{4-6}$. Simulations to-date indicate that the effect of additionally available N (affecting the carbon-climate feedback) may be small compared with the overall reduced CO$_2$-fertilisation response (the carbon-concentration feedback) in models that include C-N interactions$^{4,6}$. 
Panel c: Elevated temperatures stimulate CO$_2$ and methane emissions from global wetlands and northern permafrost soils. Enhanced NPP (e.g., in response to CO$_2$ fertilisation) can also enhance CH$_4$ emissions as more substrate is provided for microbial processes. Both increase the positive carbon-climate feedback by enhanced atmospheric methane or enhanced CO$_2$ (see panel a) concentration$^{12,49-51}$. An enhanced methane burden in the atmosphere also acts upon prolonging its own lifetime, by reducing levels of its chief oxidant OH.

(II) Anthropogenic changes in atmospheric composition alter BGC of a radiatively active compound.

Panel d: Increased tropospheric O$_3$ burden from anthropogenic pollution reduces plant photosynthesis due to its phytotoxic effects. This reduces the global terrestrial net carbon sink, resulting in accelerating CO$_2$ concentration in the atmosphere, and leading to a more positive carbon-concentration feedback$^{11}$. Enhanced CO$_2$ concentration will limit the O$_3$ phytotoxic effects to some degree as plants can operate at lower stomata opening$^{11}$. Decreased stomata conductance ($g_s$) will reduce O$_3$ deposition to canopies (ignoring, for simplicity, the likely increased canopy leaf area index at higher CO$_2$) and hence increase O$_3$ concentration. Furthermore, enhanced atmospheric CO$_2$ levels and warmer temperatures will feed back on levels of tropospheric O$_3$ by affecting BVOC and soil NO$_x$ emissions (see panel e).

(III) Climate change alters BGC of substances that in turn affect atmospheric composition of other, climatically active compounds.

Panel e: Emissions of biogenic volatile organic compounds from vegetation are stimulated by warmer temperature and enhanced NPP. In anthropogenically polluted (high NO$_x$) environment this fosters tropospheric ozone production$^{16-19}$. Stimulation of natural NO$_x$ emissions from soil by warmer temperatures will also need to be taken into account for O$_3$ formation. Models suggest that these processes result in enhanced O$_3$ levels, which in turn affect the climate, constituting a direct
climate feedback, also via the interactions plotted in (d). A further positive climate feedback of enhanced BVOC emissions is via increased methane lifetime and hence concentration, since BVOC and methane have the same atmospheric sink in their reaction with the hydroxyl radical, OH$^{19,52}$.

At least in case of isoprene, the most abundant BVOC, increasing atmospheric CO$_2$ concentration leads to a decline in leaf production. The competing effects of temperature and NPP stimulation vs. CO$_2$ inhibition on BVOC-O$_3$ interactions are not yet resolved; one study indicated on global scale little effect on overall burden (indicated as "0" along the arrows in panel (E)), but with large geographic variation$^{53}$.

Panel f: BVOC emissions are the chief precursors for biogenic secondary aerosol (SOA) growth, and determine the SOA background even in the anthropogenic atmosphere$^{29,54-55}$. SOA affect climate mainly via scattering and absorption of light (direct aerosol effect, a negative temperature feedback)$^{56}$, and by enhancing cloud albedo$^{57}$. Changes in radiation, for instance larger proportion of diffuse radiation could enhance GPP and NPP$^{58}$ (not shown). SOA are also efficient cloud condensation nuclei, with larger number concentration possibly leading to less precipitative clouds$^{59}$. Possible offsets between stimulation of BVOC emissions by warming and inhibition by increasing CO$_2$ represent a chief source of uncertainty for the magnitude of the proposed feedback loops. Faster chemistry due to elevated O$_3$ levels also increases SOA (not shown).

Citations


SUPPLEMENTARY INFORMATION

doi: 10.1038/ngeo905


