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1 **Drought-induced reduction in temperature dependence of respiration**
2 **decelerates net carbon loss with autumn warming in northern ecosystems**

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15 **Boreal and arctic ecosystems are highly sensitive to climate change, with the northern high-**
16 **latitude region warming faster than the global average (IPCC, 2013). Most previous studies**
17 **on the response of the terrestrial carbon cycle to warming have focused on the net carbon**
18 **uptake period (Lafleur et al., 2007; Richardson et al., 2009; Piao et al., 2017), while much less**
19 **attention was paid on the dormant season, during which net carbon release occurs.**
20 **Understanding how net carbon exchanges from the dormant season respond to warming is,**
21 **however, equally crucial for forecasting ecosystem–carbon cycle feedbacks. Here, we present**
22 **findings on the long-term effects of climate change on high-latitude ecosystem carbon cycle**
23 **during the dormant season from the atmospheric CO₂ concentration record (Point Barrow,**
24 **Alaska). We show that over the full study period (1974–2014), warming has significantly**
25 **boosted autumn net carbon loss and advanced the CO₂ sink-source transition date, in line**
26 **with previous analyses (Piao et al., 2008). However, in the second half of the study period, the**
27 **atmospheric CO₂ record indicates no correlation between autumn net carbon loss and**
28 **warming, which is further supported by analyses of net biome production from two different**
29 **atmospheric inversion systems. Based on multiple sources of satellite-based productivity data,**
30 **a suite of state-of-the-art ecosystem models and an atmospheric transport model, we further**
31 **suggest that this deceleration of carbon losses with warming can be attributed to the loss of**
32 **temperature dependency in respiration due to the soil moisture reduction, instead of**
33 **changing temperature-productivity relationship, and changes in atmospheric transport,**
34 **fossil fuel emissions and air-sea CO₂ exchanges. Our findings suggest that a warming climate**
35 **does not necessarily result in a higher autumn CO₂ release, which offsets recently reported**
36 **warming-induced loss of net carbon uptake during spring and summer seasons (Piao et al.,**
37 **2017; Peñuelas et al., 2017) and therefore provide a negative feedback to climatic warming.**

38 The northern land region that includes the tundra and boreal forests is acknowledged to be an
39 important component of the global carbon cycle, accounting for a considerable land-based sink for
40 atmospheric CO₂ (McGuire et al., 2009; Pan et al., 2011). There is a wide recognition that climate
41 change is having and will continue to have fundamental impacts on northern ecosystem carbon
42 cycling and in turn on variations in atmospheric CO₂ (Beer et al., 2010; Keenan et al., 2014;
43 McGuire et al., 2009; Heimann and Reichstein, 2008; Cox et al., 2000; Fiedlingstein et al., 2001;
44 McGuire et al., 2009; Ahlström et al., 2012). Although studies of ecosystem responses to warming
45 have mainly focused on spring and summer (Guerlet et al., 2013; Keenan et al., 2014; Piao et al.,
46 2017), climate-carbon cycle interactions during the dormant season could be as crucial in
47 modulating future climate change, due to the fact that a large portion of ecosystem carbon is stored
48 in the soil (Wang et al., 2011; Commane et al., 2017), and a small fractional change in soil
49 respiration with warming might significantly affect net ecosystem production and atmospheric CO₂.
50 However, we still lack the satisfactory assessment how the overall CO₂ exchange responds to
51 climate change during the dormant season.

52
53 Multiple lines of evidence that have recently emerged indicate that the response of carbon cycling
54 to recent climate change since the late 1990s are different from the previous few decades (Piao et
55 al., 2014; Piao et al., 2017; Peñuelas et al., 2017; Ballantyne et al., 2017). It has long been assumed
56 that warming advances spring phenology and increases ecosystem carbon uptake (Keeling et al.,
57 1996; Richardson et al., 2009). Although this was valid up to the 1990s, it no longer holds because
58 of the weakening temperature control of spring net primary productivity (Piao et al., 2017).
59 Furthermore, atmospheric CO₂ concentrations suggest that warm spring and summer-induced
60 increases in annual CO₂ amplitude (the difference between the annual maximum and minimum
61 concentrations within the same year), that could reflect the strength of net carbon uptake during

62 spring and summer disappeared in the last 17 years (Peñuelas et al., 2017). These multiple
63 observational signals consistently reveal a shift in the warming effect on net carbon uptake from
64 positive to neutral or even negative in spring and summer. However, it remains unclear how the
65 effects of warming on net carbon release during the dormant season change with time. There is
66 growing consensus that ecosystem productivity shows strong acclimation to warming (Oechel et
67 al., 2000; Smith and Dukes, 2013), while respiratory flux to the atmosphere from ecosystem is
68 anticipated to increase with warming. We therefore formulate the hypothesis that warming can
69 accelerate net carbon loss during the dormant season and exacerbate negative warming impact on
70 annual carbon sequestration.

71
72 We studied this hypothesis by analyzing the relationship between indicators of net carbon release
73 inferred from atmospheric CO₂ and temperature during the dormant season, and its temporal
74 change over the period 1974–2014. We calculated partial correlation coefficient between net
75 carbon release during the dormant season (defined as the change of CO₂ concentration from
76 September to November for autumn and from December to next April for winter at Point Barrow,
77 Alaska) and temperature in boreal and arctic ecosystems north of 50°N (through removing
78 statistical influence of precipitation and cloudiness variations, as detailed in Methods). There is a
79 tight relationship between autumn net carbon release (ACR) and temperature on the inter-annual
80 timescale over the period 1974–2014 ($R_{ACR-T} = -0.39$, $P < 0.05$) (Figure S1), confirming that
81 warming-induced increase in autumn respiration dominated over autumn photosynthetic gains
82 (Piao et al., 2008; Miller, 2008). Unexpectedly, R_{ACR-T} changed from -0.62 ($P < 0.01$) during
83 1974–1996 to -0.05 ($P = 0.85$) during 1997–2014, which runs counter to our proposed hypothesis
84 that the negative temperature impact on annual carbon sequestration would recently become much

85 more pronounced. The observed diminished correlation between mean autumn temperature and
86 ACR implies smaller land carbon release and reduced atmospheric CO₂ growth between September
87 and November during warmer years. The sensitivity of ACR to autumn temperature (γ_{ACR-T}) shifts
88 from -1.09 ppm K⁻¹ during the earlier period to -0.11 ppm K⁻¹ during later period (Figure S2), with
89 change in magnitude of 0.98 ppm K⁻¹, indicating a change in sensitivity of about 2.09 gigatonnes
90 of carbon per year per K when calculating based on a conversion factor of 2.14 GtC ppm⁻¹ (IPCC,
91 2013). This diminished negative temperature effect on autumn carbon cycle is also detected in the
92 upward zero-crossing date of CO₂ (defined as the day when detrended seasonal CO₂ crosses the
93 zero line from the negative to positive value). In the earlier period warmer years implied earlier
94 crossing dates ($R = -0.66$, $P < 0.01$), while in the later period no correlation remained ($R = -0.02$,
95 $P = 0.95$) (Figure S3). In contrast to autumn, in winter no temperature response of ACR was
96 detected, neither in the earlier period ($R = 0.11$, $P = 0.63$), nor in the later period ($R = -0.04$, $P =$
97 0.89) (Figure 1c).

98
99 To test the robustness of the observed decelerated loss of warming impact on autumn ACR, we
100 performed the following additional analyses: (1) we defined autumn as the period from September
101 1st to the date when detrended seasonal CO₂ crosses the zero line from the negative to positive
102 value (Figure S4), (2) we used another climate dataset (WFDEI, see Methods, Figure S5) and (3)
103 we used CO₂ concentration records from weekly *in situ* measurements and flask samples (see
104 Methods, Figure S6). All of these analyses confirmed that autumn warming no longer accelerates
105 autumn net carbon release in the latest period. In a consistent manner, we also analyzed the
106 temporal change in temperature dependence of net biome production (NBP) from two different
107 atmospheric inversion systems. Consistent with the atmospheric CO₂ analyses, NBP from the Jena
108 CarboScope inversion system also indicated a non-significant temperature impact on autumn NBP

109 over boreal and arctic ecosystems north of 50°N during 1997–2011 ($R = -0.44 \pm 0.13$), in contrast
110 to the significant temperature effect found during 1982–1996 ($R = -0.74 \pm 0.05$) (Figure 2a).
111 Similar results were also found if NBP from MACC inversion system was considered (1982–1996:
112 $R = -0.65 \pm 0.08$; 1997–2011: $R = -0.16 \pm 0.15$, Figure S7). Besides land ecosystems, atmospheric
113 CO₂ variation also harbors signals from changes in atmospheric transport, air-sea CO₂ exchanges
114 and fossil fuel emissions. We therefore assessed their potential contributions to the change in R_{ACR} .
115 T using atmospheric transport simulations based upon atmospheric transport model from the
116 Laboratoire de Météorologie Dynamique (LMDz) (Hourdin et al., 2006) (see Methods), and found
117 that their decadal changes would not contribute to the observed diminished temperature control on
118 ACR (Figure S8).

119
120 Which terrestrial carbon cycle processes caused the diminished negative temperature control on
121 the autumn carbon cycle in the north? This diminished effect could be only explained by an
122 enhanced temperature reliance of carbon uptake through vegetation photosynthesis and/or a
123 reduced temperature dependence of carbon loss through respiration. Analysis of satellite-based
124 vegetation index (GIMMS NDVI) (Tucker et al., 2005) as a proxy for vegetation production
125 showed that autumn NDVI is marginally significantly correlated with temperature in the earlier
126 period ($R = 0.50 \pm 0.15$), but became decoupled from temperature in the later period ($R = -0.37 \pm$
127 0.14) (Figure S9). This weakened temperature dependence of vegetation production was,
128 nonetheless, also evident when considering satellite-based estimates of net primary productivity
129 (NPP) (Smith et al., 2016, Figure 2b), or satellite-independent estimates of gross primary
130 productivity (GPP) up-scaled from eddy flux towers (Jung et al., 2009)(Figure S10), thereby ruling
131 out its possibility in explaining the diminished temperature effect on ACR. For example, R_{NPP-T}

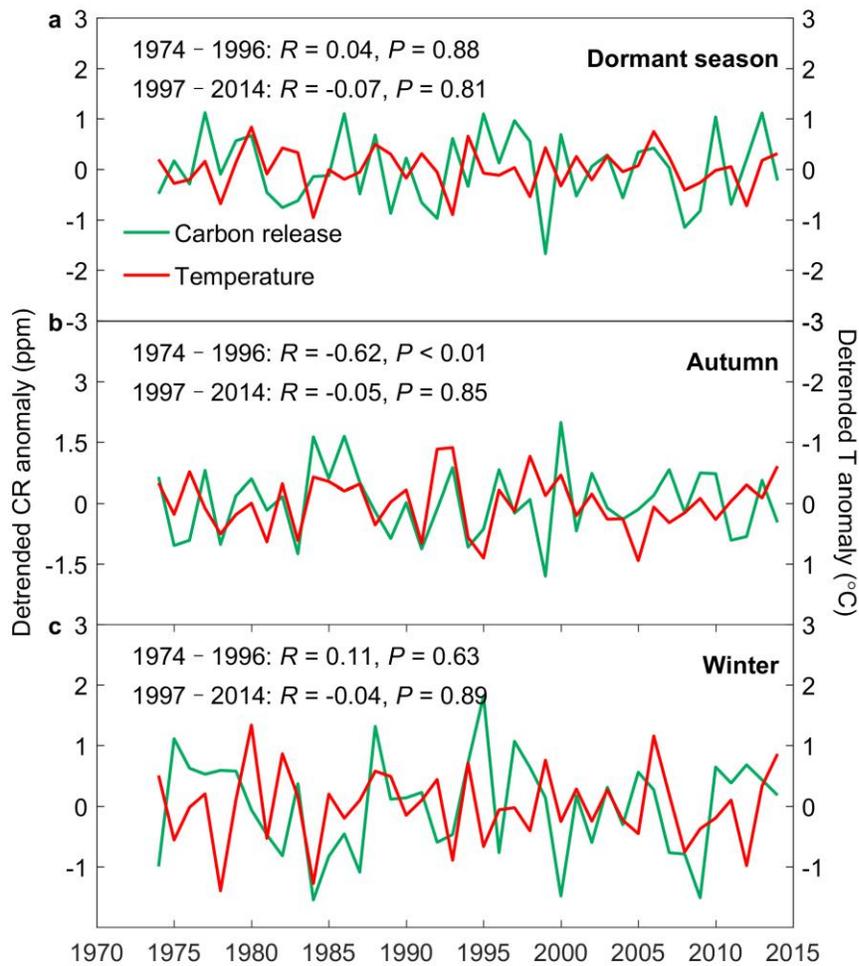
132 and R_{GPP-T} decreased from 0.82 ± 0.06 and 0.88 ± 0.04 in the earlier period to 0.41 ± 0.23 and 0.43
133 ± 0.14 in the later period, respectively.

134
135 The heterotrophic respiration (HR), computed as the difference between NBP from Jena
136 CarboScope (or MACC) dataset and satellite-derived NPP, has significant partial correlations with
137 autumn temperature in the earlier period (Jena: $R_{HR-T} = 0.90 \pm 0.03$; MACC: $R_{HR-T} = 0.79 \pm 0.06$)
138 but not in the later period (Jena: $R_{HR-T} = 0.48 \pm 0.14$; MACC: $R_{HR-T} = 0.28 \pm 0.15$) (Figure 2c,
139 Figure S7b). Furthermore, we also analyzed simulated HR from eight models participating in the
140 historical climate carbon cycle model intercomparison project (TRENDY, see Methods), and found
141 that the strong HR-temperature correlation in the earlier period also became weak and non-
142 significant during the later period across almost all of the models (Figure S11 and S12). Therefore,
143 we conclude that the diminished negative temperature effect on autumn carbon cycle during the
144 latest period is most likely due to diminished temperature dependence of respiratory losses. To
145 diagnose the potential mechanism responsible for the decrease in R_{HR-T} , we studied decadal changes
146 in simulated soil moisture content from TRENDY models, and found a widespread reduction in
147 soil moisture particularly over North America and Siberia, which was spatially coherent with the
148 decline in R_{HR-T} , suggesting the plausibility of a potential soil water effect (Figure S13).

149
150 We provide the evidence that autumnal warming no longer accelerates net carbon losses and
151 advance the end of the carbon uptake period in boreal and arctic ecosystem as previously suggested
152 (Piao et al., 2008; Ueyama et al., 2014), primarily through reducing positive decomposition
153 responses to warming most likely due to a soil moisture shortage. The autumnal finding reveals the
154 similar temperature-dependent shift in carbon cycle over the last 3 decades that is also found to
155 occur in the main growing season (Peñuelas et al., 2017; Piao et al., 2017), suggesting a changing

156 paradigm for temperature control over ecosystem carbon cycling. However, the outcomes of these
157 shifts on net CO₂ exchanges are not consistent in the direction of their effect on atmospheric pCO₂
158 and would thus partly compensate for each other. The autumnal respiratory acclimation has an
159 ameliorating impact on net CO₂ losses with rising temperatures, which could offset the negative
160 warming impact on net CO₂ uptake during the active growing season (Peñuelas et al., 2017; Piao
161 et al., 2017). It is therefore premature to conclude that the impact of temperature on annual carbon
162 cycle has fundamentally shifted towards the negative state, and highlights the importance of
163 incorporating how net carbon losses change with temperature during the dormant period in fully
164 understanding temperature impacts on net carbon uptake. Additional studies are still needed to
165 quantify whether these two opposing effects on carbon cycle will effectively neutralize each other,
166 particularly for arctic and boreal ecosystems where the majority of permafrost soil carbon is stored
167 and increasing old soil carbon will be respired to the atmosphere as a result of warming-induced
168 permafrost thaw (Schuur et al., 2015; Pries et al., 2016; Koven et al., 2011).

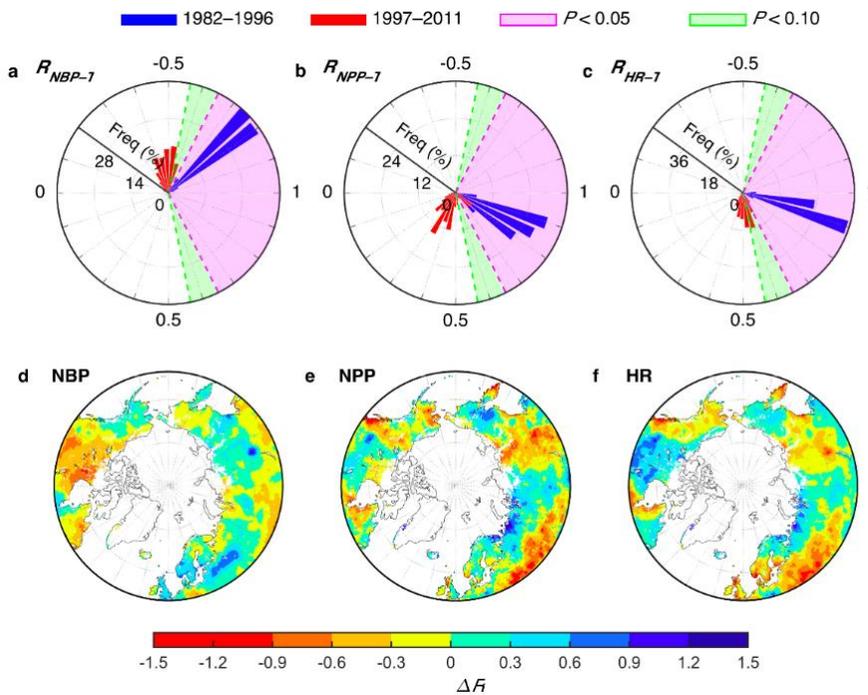
169 **Figure 1. Temperature control on net carbon release during the dormant season.** Here we
 170 define the dormant season as the period from September to next April (**a**), which consists of autumn
 171 (September to November, **b**) and winter (December to next April, **c**). The lines are time series of
 172 the detrended anomaly of net carbon release (green) and mean temperature across land ecosystems
 173 north of 50°N (red).



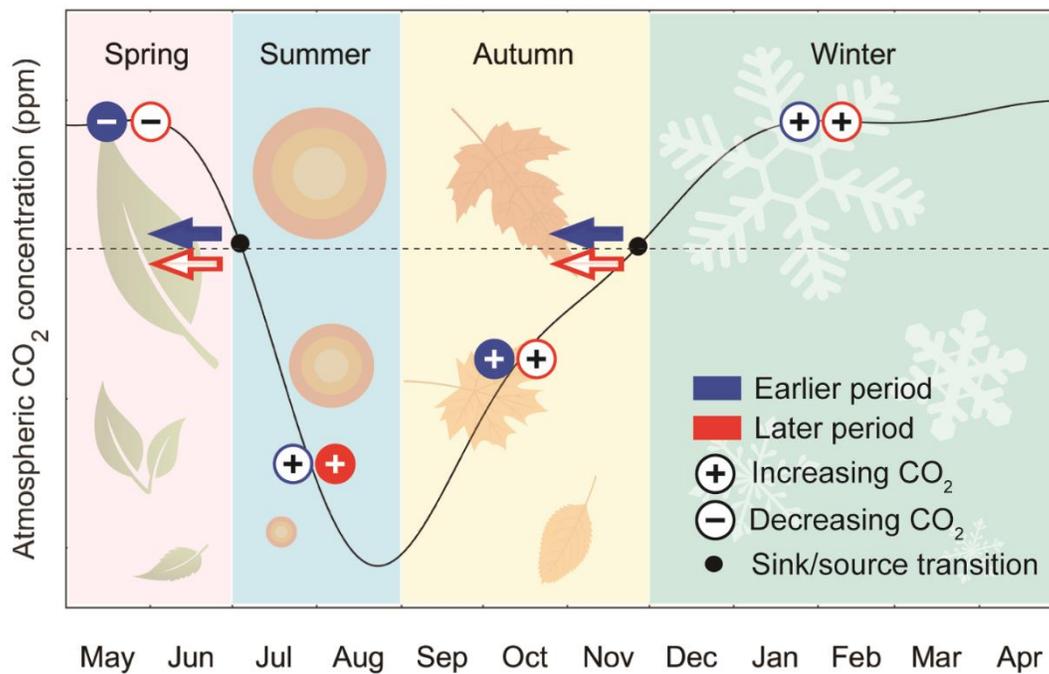
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176 **Figure 2. The relationship between ecosystem carbon fluxes and temperature in autumn.** (a-
 177 c), the frequency distribution of the partial correlation coefficient of net biome production (NBP),
 178 net primary productivity (NPP), and heterotrophic respiration (HR) with average temperature
 179 during September to November across land ecosystem north of 50°N, whilst controlling for
 180 precipitation and cloudiness during the earlier period (1982–1996, blue bar) and later period
 181 (1997–2011, red bar), respectively. For each period, we randomly selected 12 years to generate the
 182 frequency distribution of partial correlation coefficient. The shade illustrates the significance level
 183 at $P < 0.05$ (magenta) and $P < 0.10$ (green), respectively. (d-f), the spatial distribution of the
 184 changes in the partial correlation coefficient of NBP, NPP and HR with temperature for the two
 185 periods, respectively. Here NBP is estimated from Jena CarboScope inversion system, the NPP is
 186 estimated based on GIMMS NDVI, and HR was calculated as the difference between NBP and
 187 NPP.



189 **Figure 3. Schematic of the effect of warming on seasonal CO₂ uptake and release.** In spring,
 190 warming advances the source-to-sink transition date and increases CO₂ uptake that decreases
 191 atmosphere CO₂ (Keeling et al., 1996), but which disappeared in the later period (1996–2012)
 192 (Piao et al., 2017). In summer, the effect of warming on net CO₂ uptake became significantly
 193 negative in the later period that increases atmosphere CO₂ (Peñuelas et al., 2017). On the contrary,
 194 the widely recognized autumn warming-induced advancement in sink-to-source transition date and
 195 acceleration in net CO₂ release (Piao et al., 2008) became diminished in the later period, which
 196 could decrease build-up of atmospheric CO₂. In contrast, the temperature effect on winter net CO₂
 197 release is not significant during both periods.



198

199 **Materials and Methods**

200 **Atmospheric CO₂ concentration**

201 The CO₂ concentration records from Point Barrow (71°N, Alaska) cover the period from 1974 to
202 2014, and are derived from the National Oceanic and Atmospheric Administration (NOAA) Earth
203 System Research Laboratory (Thoning et al., 2014). The CO₂ concentration time series consist of
204 three types of signals: the long-term trend, the short-term variations, and the seasonal cycle. We
205 performed the following procedure to obtain detrended seasonal cycle of CO₂ concentration. First,
206 we fitted the daily CO₂ records using a function consisting of four harmonics and a quadratic
207 polynomial to separate the seasonal cycle from the long-term increasing trend (Thoning et al, 1989)
208 and obtain the residuals from this function fit. Second, we used a 1.5 month (or 1 month, see Figure
209 S6) full-width half-maximum value (FWHM) averaging filter to remove the short-term variations
210 from the residuals and get a smoothed curve by adding the filtered residuals to the fitted function
211 in the first step. We also applied a 390-day FWHM averaging filter to derived residuals from the
212 first step and added the filtered residuals to the fitted long-term trend from quadratic polynomial
213 to obtain a de-seasonalized long-term trend. Finally, we calculated the difference between the
214 smoothed curve and the de-seasonalized long-term trend as the detrended seasonal CO₂
215 concentration. As outlier records have strong influence on the fitting process, we repeatedly fitted
216 the CO₂ time series as described in the first step and discarded records lying outside five times of
217 standard deviation of the residuals until no outliers were found (Harris et al, 2000).

218

219 We define the dormant season as the period from September to next April and calculated the
220 changes in detrended CO₂ concentration during this period (CO₂ concentration in the last week of
221 April in next year minus that in the first week of September) as the dormant season net carbon

222 release (CR). We also separate the dormant season into autumn (September to November) and
223 winter (December to next April) and calculated autumn carbon release (ACR) and winter carbon
224 release (WCR) as the changes of CO₂ concentration in autumn and winter, respectively. In addition,
225 the mean date when detrended seasonal CO₂ crosses zero line from the negative to positive value
226 is around the 317th day of the year (DOY) during the period from 1974 to 2014. To test the
227 robustness of the analysis, we also defined autumn as the period from the first day of September to
228 DOY 317 and defined winter as the period from DOY 318 to the last day of April in next year and
229 calculated ACR and WCR accordingly. Furthermore, we also calculate CR from the weekly
230 atmospheric CO₂ concentration from the NOAA Earth System Research Laboratory at Barrow.

231

232 **Climate dataset**

233 We used the monthly climate dataset from the Climate Research Unit, University of East Anglia
234 (CRU TS4.0 dataset) (Mitchell et al, 2005) in this study. This dataset covers the period from 1901
235 to 2015, with a spatial resolution of 0.5°×0.5°. We selected mean temperature, precipitation and
236 cloud cover for the analysis. We also used another climate dataset, which applied the WATER and
237 global Change (WATCH) Forcing Data to the ERA-Interim dataset ([http://www.eu-
238 watch.org/gfx_content/documents/README-WFDEL.pdf](http://www.eu-watch.org/gfx_content/documents/README-WFDEL.pdf)) for the analysis and obtained similar
239 results (Figure S5).

240

241 **Vegetation production datasets**

242 We used the Normalized Difference Vegetation Index (NDVI) retrieved from the third-generation
243 of the Advanced Very High Resolution Radiometer (AVHRR) developed by the Global Inventory
244 Modeling and Mapping Studies (GIMMS) group (version 3g.v0, available at

245 <https://ecocast.arc.nasa.gov/data/pub/gimms/3g.v0>) as a proxy for vegetation activity (Tucker et al,
246 2005). The GIMMS NDVI dataset covers the period from 1982 to 2013, with a spatial resolution
247 of $0.083^{\circ} \times 0.083^{\circ}$. We also used two vegetation production data: the monthly GIMMS net primary
248 production (NPP) dataset (Smith et al, 2016), and the gross primary productivity (GPP) up-scaled
249 from eddy flux towers using multi-tree ensemble approach (Jung et al, 2009).

250

251 **Atmospheric CO₂ inversion data**

252 We gathered two atmosphere CO₂ inversion products to investigate the response of terrestrial
253 carbon fluxes to warming. We used monthly net biome production (NBP) from the Jena
254 CarboScope (<http://www.bgc-jena.mpg.de/CarboScope/>, version s81_v3.8) for the period from
255 1982 to 2011, with a spatial resolution of 3.75° latitude $\times 5^{\circ}$ longitude. The monthly net biome
256 production (NBP) from the Monitoring Atmospheric Composition and Climate (Chevallier et al,
257 2005) (MACC, version v14r2, <http://copernicus-atmosphere.eu/>) between 1979 and 2011 was also
258 used for the analysis. We calculated the heterotrophic respiration as the difference between
259 inversed NBP and satellite-based NPP.

260

261 **Terrestrial ecosystem models**

262 Simulation results of eight models from a historical climate carbon cycle model inter-comparison
263 project (Trendy) were used in this study. These models are Community land Model Version 4.5
264 (CLM4.5), the Integrated Science Assessment Model (ISAM), the Joint UK Land Environment
265 Simulator (JULES), Lund-Potsdam-Jena DGVM (LPJ), Lund-Postam-Jena General Ecosystem
266 Simulator (LPJ-GUESS), the Land surface Processes and eXchanges (LPX), the Organizing
267 Carbon and Hydrology In Dynamic Ecosystems (ORCHIDEE), and the Vegetation Integrative
268 Simulator for Trace gases (VISIT). All the models used forcing data from CRUNCEP dataset, and

269 the simulation setup follow the standard protocol described in the inter-comparison project
270 (http://dgvn.ceh.ac.uk/files/Trendy_protocol%20Nov2011_0.pdf). Here we used the S2
271 simulations, which consider the effect of climate change and rising CO₂ concentration on
272 ecosystem carbon fluxes.

273

274 **Effects of atmospheric transport, air-sea CO₂ exchanges and fossil fuel emission on the**
275 **change in autumn net CO₂ release**

276 To investigate the effects of atmospheric transport, air-sea CO₂ exchanges and fossil fuel emission
277 on the change in autumn net carbon release, we assessed the impact of year-to-year variations in
278 atmospheric transport, air-sea CO₂ exchange and fossil fuel emission on the observed changes on
279 autumn net carbon release between the early period (1979–1996) and the later period (1997–2012)
280 using atmospheric transport simulations. We used LMDz4, a 3D atmospheric tracer transport
281 model from the Laboratoire de Météorologie Dynamique (Hourdin et al., 2006), nudged with
282 ECMWF winds. As boundary conditions for transport simulations, we use land carbon fluxes over
283 1979–2012 from the land surface model ORCHIDEE (Krinner et al., 2005) that is driven by
284 observed atmospheric CO₂ concentration and historical climate forcing from the CRU-NCEPv4
285 climate variables at 6-h resolution (Viovy and Ciais, 2014). For air–sea CO₂ exchanges, we use
286 simulations from a biogeochemical model PlankTOM5 combined with a global ocean general
287 circulation model NEMO (NEMO-PlankTOM5) that is forced by inputs of ions and compounds
288 from river, sediment and dust for the PlankTOM5 model, and daily wind and precipitation from
289 the NCEP reanalysis for the NEMO model (Buitenhuis et al., 2010). For fossil fuel CO₂ emissions,
290 the monthly global time series was derived from the Carbon Dioxide Information Analysis Center
291 (CDIAC) website (<http://cdiac.esd.ornl.gov>) (Andres et al,2011).

292
293 To assess whether changes in atmospheric transport can influence the observed change in ACR,
294 we perform the transport modeling experiment in which land and air–sea CO₂ exchanges are fixed
295 at the year 1979 but the atmospheric transport allows to be varying according to ECMWF wind
296 fields (refer to WCC hereafter). For air–sea CO₂ exchange, we conduct the modeling experiment
297 where the atmospheric transport and land carbon fluxes are fixed at year 1979 but air–sea CO₂
298 exchanges vary according to simulations from NEMO-PlankTOM5 (WAC simulation). To assess
299 the effect from fossil fuel emission, we conducted the modeling experiment where the land and air-
300 sea CO₂ exchanges fixed at the year 1979, but transport the year-to-year varying fossil fuel
301 emission (WCF simulation).

302
303 **Analysis**
304 We performed partial correlation analysis between net carbon release during the dormant season
305 (autumn and winter) with temperature whilst statistically controlling for precipitation and cloud
306 cover (R_{CR-T} , R_{ACR-T} , and R_{WCR-T}). The climate variables are averaged over the region north of 50°N,
307 and we only considered the pixels where the annual NDVI greater than 0.1. The partial correlation
308 analysis was performed for the earlier period (1974-1996) and later period (1997-2014)
309 respectively. All variables are detrended before the partial correlation analysis. For a more robust
310 analysis, we also performed the partial correlation analysis through randomly selecting 12 years
311 from the time series among the corresponding period to generate a frequency distribution of the
312 partial correlation coefficient. We also conducted a two-sample *t*-test to determine whether the
313 partial correlation coefficient is statistically significant. To test if the shift of R_{ACR-T} is influenced
314 by atmospheric transport, we calculated ACR from the WCC simulation, which all factors except
315 wind field are fixed to year 1979, to denote the effect from atmospheric transport. To test if the

316 shift of R_{ACR-T} is influenced by the transport of air-sea CO_2 and fossil fuel emission, we calculated
317 the air-sea CO_2 and fossil fuel induced ACR by calculating the difference between the WAC (WCF)
318 and the WCC simulation. Then we conducted the partial correlation analysis on the WAC (WCF)
319 induced ACR. To investigate the driver of the shift of R_{ACR-T} , we also performed the same analysis
320 to the satellite-derived NDVI (R_{NDVI-T}), the satellite-based NPP (R_{NPP-T}), the flux-tower based GPP
321 (R_{GPP-T}), the inversed NBP (R_{NBP-T}) and the HR calculated from NBP and NPP (R_{HR-T}).

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