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1 **Title:** Centennial-scale variations in the carbon cycle enhanced by high obliquity

2

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20

21 **Abstract:** Centennial-scale increases of atmospheric carbon dioxide, known as carbon dioxide
22 jumps, are identified during deglacial, glacial and interglacial periods and linked to the
23 Northern Hemisphere abrupt climate variations. However, the limited number of identified
24 carbon dioxide jumps prevents investigating the role of orbital background conditions on the
25 different components of the global carbon cycle that may lead to such rapid atmospheric
26 carbon dioxide releases. Here, we present a high-resolution carbon dioxide record measured
27 on an Antarctic ice core between 260 and 190 thousand years ago which reveals seven
28 additional carbon dioxide jumps. 18 of the 22 jumps identified over the past 500 thousand
29 years occurred under a context of high obliquity. Simulations performed with an Earth system
30 model of intermediate complexity point toward both the Southern Ocean and the continental
31 biosphere as the two main carbon sources during carbon dioxide jumps connected to Heinrich
32 ice rafting events. Notably, the continental biosphere appears as the obliquity-dependent

33 carbon dioxide source for these abrupt events. We demonstrate that the orbital-scale external
34 forcing directly impacts past abrupt atmospheric carbon dioxide changes.

35 **Main text:** Understanding the mechanisms of natural centennial-scale changes in the carbon
36 cycle is a major challenge under the ongoing anthropogenic carbon dioxide (CO₂)-driven
37 climate change. Recent efforts to produce high-resolution atmospheric CO₂ records from
38 Antarctic ice cores demonstrated the pervasive character of centennial-scale atmospheric CO₂
39 increases (Carbon Dioxide Jumps, CDJs) in the global climate system over the last 500 ka
40 (thousand years before present)¹. These CDJs consist of an atmospheric CO₂ increase from ~5
41 ppm up to ~ 16 ppm, that occurs in a few decades to centuries at estimated rates of up to 26
42 ppm/century in the atmosphere¹. The CDJs are identified through thresholds applied on
43 detrended ice core CO₂ records (amplitude larger than 5 ppm and growth rate equal or above
44 1.5 ppm/century after removing the orbital trend)¹. Hence, they only represent the strongest
45 centennial-scale atmospheric CO₂ increases. Two types of Northern Hemisphere (NH) climatic
46 events are associated with the CDJs: (i) the Dansgaard-Oeschger (DO) events which are
47 identified during the last glacial period in Greenland^{2,3} and North Atlantic⁴ archives. Each DO
48 event is characterized by an abrupt transition from cold - *stadial* - to milder - *interstadial*-
49 conditions associated with a Greenland surface temperature increase of 5 to 16°C in less than
50 a century⁵. We stress that a CDJ with a minimum amplitude of 5 ppm cannot be detected for
51 all DO warmings; (ii) the maximum intensity of Heinrich events (HE) within the Heinrich
52 stadials (HS), i.e. cold stadial periods in glacial times during which Heinrich iceberg discharge
53 events occurred, as reflected in peaks of Ice Rafted Debris (IRD) in North Atlantic sediment
54 cores⁶. The CDJs occurring at the onset of a DO event (hereafter CDJ+) ¹ are associated with
55 an increase in atmospheric methane (CH₄) concentrations larger than 50 ppb⁷. The CDJs
56 occurring within a HS (hereafter CDJs-) ¹ are associated with a short and smaller CH₄ increase
57 ¹ happening within the HS and likely when maximum iceberg discharge led to further reduction
58 of the AMOC (Atlantic Meridional Overturning Circulation). Proposed hypotheses to explain
59 how these rapid atmospheric CO₂ increases relate to DO events or HE involve both terrestrial
60 ^{8,9} and marine CO₂ sources^{10,11}, but overall the underlying mechanisms remain poorly
61 understood.

62 Here, we argue that the orbital-scale climatic background conditions modulate the intensity
63 of the centennial-scale carbon cycle response to the Northern Hemisphere climatic

64 perturbations. To test this hypothesis, we extend the time interval covered by high-resolution
65 atmospheric CO₂ reconstructions to allow the identification of additional potential CDJs under
66 varying orbital configurations. Then, we investigate the impact of orbital-scale changes in the
67 climatic background on centennial-scale CO₂ increases through simulations performed with
68 an Earth system model of intermediate complexity.

69 **Sub-millennial-scale atmospheric CO₂ record**

70 We measured the CO₂ concentrations on the air trapped in 203 samples from the Antarctic
71 EPICA Dome C (EDC) ice core between 260 and 190 ka using a dry-extraction system coupled
72 to a gas chromatograph¹²; (see methods). Our CO₂ record is in agreement with published data
73 ¹³ while the mean temporal resolution is higher, 340 years on average (Fig. 1, Supplementary
74 Figure 1). We identify seven new CDJs using the detection methodology described in ref. ¹. In
75 summary, we first remove the orbital-scale (i.e. deglaciation and glaciation) and millennial-
76 scale (i.e., Antarctic Isotope Maximum events) imprint in the CO₂ record using cubic smoothing
77 splines ¹⁴. Second, we calculate the rate of change of the detrended CO₂ record and third, we
78 apply thresholds of 1.5 ppm/century for the minimum rate of increase and 5 ppm for the
79 minimum amplitude from ref. ¹ to identify CDJs (see methods; Supplementary Figures 2,3). In
80 our record, we identify two CDJs+ associated with atmospheric CH₄ increases higher than 50
81 ppb in less than 300 years (CDJ+7e and CDJ+7a), three CDJs- that are not associated with a
82 detectable increase in atmospheric CH₄ in the ice core record and that can be linked to a IRD
83 peak in the oceanic record (CDJ-8a.1, CDJ-8a.2, CDJ-8a.3), and two CDJs that are not associated
84 with either of these two markers (CDJ 7c, CDJ 7b) (Fig. 1, Supplementary Figures 3,4). Two
85 CDJs occurred during interglacial conditions¹⁵ (CDJ 7b, CDJ+7a) when Northern Hemisphere
86 ice sheets had shrunk to a size approaching the Holocene value, confirming the occurrence of
87 CDJs during periods of relatively small global ice sheet volume ¹ (Fig. 1, Supplementary Figure
88 5). Hence, combining published and newly-described events we highlight 22 CDJs occurring
89 under a wide range of climatic and orbital boundary conditions over the past 500 ka (Fig. 2,
90 Supplementary Figure 6, Supplementary Table 1).

91 **Carbon cycle variability and obliquity state**

92 The sources and mechanisms behind centennial-scale releases of atmospheric CO₂ have been
93 investigated previously based on carbon isotope measurements on Antarctic ice and carbon

94 cycle modelling^{8,16}. Two main hypotheses have been discussed: (i) the role of a terrestrial
95 carbon source due to a southward shift in the intertropical convergence zone^{8,17} and (ii) an
96 increase of upwelling activity in the Southern Ocean and/or the North Pacific Ocean possibly
97 due to enhanced intensity or location relative to the sea-ice edge of winds^{10,18}. However, the
98 role played by long-term forcing on the centennial-scale CO₂ variability has not yet been
99 assessed.

100 Here we investigate the potential role of the long-term forcing, e.g. the global ice volume and
101 the orbital forcing, on the occurrence of CDJs. We show that 18 of the 22 CDJs occurred at
102 times above the average obliquity value (23.33°) of the 12 obliquity cycles of the last 500 ka
103 (Fig. 2; Supplementary Figure 5). We note that seven of the eight obliquity maxima between
104 416 and 9 ka covered by high-resolution CO₂ and CH₄ data that allows the identification of CDJ
105 occurrences show at least one CDJ. Reversely, only two out of the seven obliquity minima
106 covered by high-resolution CO₂ data are linked to a CDJ (Fig. 2). We investigate the hypothesis
107 of a coincidental origin of this specific pattern of CDJs relative to obliquity changes performing
108 a statistical chi-square test (see Methods). According to our analysis, the null hypothesis that
109 centennial-scale release of CO₂ is independent of the obliquity state can be rejected with 90%
110 confidence ($p < 0.1$). Reversely, the occurrence of CDJs is not significantly coupled to the
111 changes in climatic precession, eccentricity, and global ice volume (Supplementary Figure 5;
112 Supplementary Table 2). Note that our results are independent from the ice-core gas
113 chronology used to display the CO₂ record^{19–22} and the choice of orbital solutions^{23,24}
114 (Supplementary Figure 7, Supplementary Table 3). In addition, while our results are obtained
115 using the threshold detection values from ref.¹, sensitivity tests show that obliquity
116 dependence of the CDJ occurrences is robust regardless of the specific threshold values
117 considered (Supplementary Figure 8, Supplementary Table 4). Finally, the AMOC variability
118 itself, at the origin of the CDJs, seems not directly related to obliquity. Indeed, most of AMOC
119 variability occurs under a CO₂ concentration window of 190–225 ppm²⁵, which corresponds to
120 periods equally distributed between high (50.7%) and low (49.3%) obliquity over the past 500
121 ka.

122 **Modelling assessment of CO₂ sources during CDJs- events**

123 CDJs- are most likely associated with a complete AMOC shutdown that occurs sometime
124 within the HS¹. To study the impact of such an AMOC shutdown on the centennial-scale

125 atmospheric CO₂ response, we perform a suite of sensitivity experiments to mimic the carbon
126 cycle response to a HS under both high and low obliquity boundary conditions with the Earth
127 System Model LOVECLIM²⁶. We propose that HS 5 is the most appropriate HS to use as a case
128 study since the corresponding CDJ occurs during a period characterised by the highest
129 obliquity value (24.3°).

130 We perform four simulations under 49 ka boundary conditions, i.e., with realistic orbital
131 parameters, including an obliquity value of 24.3°, as well as an estimate of northern
132 hemispheric ice-sheet extent, topography and associated albedo²⁶ (Methods, Supplementary
133 Table 5). We also perform four simulations under the same boundary conditions except for an
134 obliquity value of 22.1°, i.e. the lowest value of the last 500 ka (Supplementary Table 5). In all
135 experiments, an idealized HE is simulated by adding 0.3 Sverdrup (Sv) of freshwater to the
136 North Atlantic (60-10°W, 50-60°N) during 1000 years, to induce an AMOC shutdown
137 (Supplementary Figure 9). The sequence of changes during HS is still debated and especially
138 the role of freshwater flux in the observed AMOC weakening^{7,27}. Nevertheless, previous
139 modelling studies have shown that the simulated climatic response to an AMOC weakening
140 was similar whether the freshwater flux had an active (hosed) or passive (unhosed) role²⁸.

141 The simulation conducted under 49 ka climatic conditions (*HighObl*) reveals a centennial-scale
142 response of the carbon cycle to the induced freshwater perturbation (Fig. 3a), characterised
143 by a ~9.5 ppm increase in atmospheric CO₂ over the first 400 years of the simulation (Fig. 3a,
144 Fig. 3c, Supplementary Table 5, Supplementary Figure 10). A similar simulation performed
145 under low obliquity (*LowObl*), only leads to a ~3.3 ppm CO₂ increase over the first 400 years
146 following the freshwater perturbation (Fig. 3c, Supplementary Table 5). In both experiments,
147 the CO₂ increase is due to a terrestrial carbon release due to climatic changes induced by the
148 AMOC weakening (Fig. 3d, Supplementary Figure 9). Surface air temperature is lower under
149 low obliquity, particularly in the Northern Hemisphere (-1.4 °C north of 20°N), thus impacting
150 the precipitation patterns (Fig. 4a). As a result, the terrestrial carbon reservoir is smaller under
151 low than high obliquity (-105 Gigatons of carbon, GtC; Fig. 4b). The decrease of precipitation
152 and the lower temperature over the Northern high-latitudes during the freshwater input thus
153 leads to a smaller terrestrial carbon release under low obliquity compared to high obliquity.

154 To investigate the role of terrestrial carbon release, two freshwater experiments are
155 performed in which carbon fluxes between the continental biosphere and the atmosphere are

156 suppressed. Under such conditions, neither the experiment performed under high obliquity
157 (*HighObl_noVeg*) nor the one performed under low obliquity (*LowObl_NoVeg*) produce a
158 centennial-scale CO₂ event (i.e. ~0 ppm CO₂ change at year 400; Fig. 3c, Supplementary Figure
159 11, Supplementary Table 5). Those results highlight that the obliquity no longer has an
160 influence on the centennial-scale CO₂ variability if the continental biosphere is muted. They
161 point to the continental biosphere as the variable source of carbon at centennial-scale for CDJ-
162 events, which appears strongly controlled by Earth's obliquity.

163 Additional experiments were conducted to investigate the impact of obliquity on the CDJ+
164 occurring at the end of the Younger Dryas (~12 ka)¹⁷ (see Methods). Our LOVECLIM
165 simulations forced with constant 12 ka boundary conditions are not able to reproduce the
166 magnitude and duration of the CDJ+ event, regardless of the obliquity forcing applied
167 (Supplementary Figures 21,22). A decrease in CO₂ solubility due to ocean warming²⁹ or
168 strengthening/poleward shift of the SHW, as suggested for the end of the YD^{30,31}, might be
169 necessary to simulate such a rapid CO₂ increase³².

170 A stronger CO₂ release from the Southern Ocean, potentially resulting from enhanced
171 Southern Ocean upwelling has been suggested as a process leading to CDJ- during Termination
172 I¹⁰. Enhanced upwelling of CO₂ rich waters in SO could result from a poleward shift of the
173 sea-ice edge¹⁰ or intensified (and/or poleward shifted) Southern Hemisphere Westerlies
174 (SHW) during HS³². The robustness of the modelled responses to such wind changes was
175 tested by conducting two additional simulations with an imposed ~ 40% increase of the SHW
176 windstress, one under 49 ka conditions (*HighObl_SHW*) and one under low obliquity
177 conditions (*LowObl_SHW*) (Supplementary Table 5). This leads to a stronger CO₂ release
178 during both high and low obliquity conditions. However, the difference in total CO₂ release
179 between high and low obliquity conditions persists and only exists when the terrestrial
180 biosphere exchanges CO₂ with the atmosphere. These results reinforce the idea that the SHW
181 play a role in centennial-scale carbon cycle increases (Fig. 3a, Supplementary Table 5).
182 However, larger CO₂ changes are simulated when terrestrial vegetation and enhanced SHW
183 (*HighObl_SHW* and *LowObl_SHW*) are considered. Our results show that the terrestrial
184 biosphere reservoir is the main source of the higher carbon release during high obliquity
185 conditions for CDJ- events due to its apparent sensitivity to the obliquity context.

186 Now, we compare the magnitude and duration of the centennial-scale atmospheric CO₂
187 variations recorded in the WAIS Divide ice core (WDC) across HS 5 (~49 ka) and those modelled
188 in the simulations that include terrestrial carbon fluxes. We look at the CO₂ difference
189 between the high- and low-obliquity simulations (ΔCO_2 , Fig. 3b). The ΔCO_2 value allows to
190 identify the difference of amplitude of the modelled centennial-scale response of atmospheric
191 CO₂ concentrations in the high-obliquity vs low-obliquity simulations. ΔCO_2 increases at
192 centennial-scale and reaches a maximum around year 400 (Fig. 3b, Supplementary Figure 10).
193 In terms of magnitude, this ΔCO_2 partially explains the recorded centennial-scale variability in
194 the WDC CO₂ record. This comparison illustrates a similar temporal structure between the
195 measured and simulated CO₂ changes. After year 1000, the WDC CO₂ changes are decoupled
196 from the variations of the simulated ΔCO_2 , revealing a change in the dynamics of the carbon
197 cycle from centennial-scale to millennial-scale most likely not related to obliquity state. While
198 this model-data comparison is limited by the modelling design, it supports our hypothesis that
199 the orbital-scale obliquity changes influence centennial-scale carbon cycle variations through
200 the terrestrial carbon reservoir.

201 **Drivers of centennial-scale atmospheric CO₂ increases**

202 Previous studies have evidenced an acidification of the Southern Ocean ^{10,11} and of the North
203 Pacific Ocean ¹⁸ synchronous with the CDJ+ occurring at the onsets of the Bolling Allerød and
204 of the Holocene. This would be consistent with an enhanced CO₂ outgassing in the Southern
205 Ocean resulting from the interplay between the location of the SHW and the sea ice edge, that
206 could have also contributed to previous CDJ+ events ¹⁰. Our modelling results show that
207 variations in SHW can act as a key driver of centennial-scale carbon cycle. However, the SHW
208 changes resulting from an AMOC shutdown could also be modulated by obliquity. Hence, we
209 cannot exclude the hypothesis that the Southern Ocean is an obliquity-dependent source of
210 CO₂ and additional simulations investigating the link between SHW and obliquity should be
211 performed with alternative models.

212 Our simulations indicate that the continental biosphere provides an additional source of
213 carbon to the atmosphere during CDJ- events, as already inferred by modelling or carbon-
214 isotope approaches ^{8,33,34} and reveals the obliquity-dependence of this carbon source at
215 centennial-scale. Especially, the region between 8°N and 40°N appears as the additional
216 source of terrestrial carbon under a high-obliquity phase (Fig. 4c). Previous studies revealed

217 an orbital- and millennial-scale control of obliquity on vegetation changes based on pollen ^{35,36}
218 and sediment records ^{37,38}, as well as through modelling approaches ³⁹. This obliquity control
219 modulates the carbon stock available to be released during centennial-scale increases through
220 the extent and type of vegetation. In the low-latitude regions, the induced southward shift of
221 the Inter Tropical Convergence Zone (ITCZ) and changes in latitudinal insolation gradients ³⁹
222 cause an obliquity-paced vegetation change especially for subtropical biomes and tree
223 coverage ^{35,38}. Depending on the area considered in the mid-latitudes regions, both precession
224 and obliquity forcing seem to be the main drivers of the Asian monsoon system, and hence of
225 the induced summer precipitation that controls the vegetation coverage in the area ³⁷. At
226 higher latitudes, the predominance of trees and shrubs over grasses and herbs is favoured
227 under high obliquity due to lower local ice extent ³⁶. Our study suggests that the expansion of
228 continental biomass in the high latitudes and the change of vegetation at low latitudes in high
229 obliquity phases could thus provide a large amount of carbon to be released during CDJ- (Fig.
230 3d). In principle, this additional carbon is also available for enhanced terrestrial carbon release
231 during CDJ+ under high obliquity conditions. With respect to the opposite latitudinal shift of
232 the ITCZ between DO and HE, a similar obliquity dependence on the terrestrial biosphere
233 carbon release is also possible. However, our model experiments performed at 12 ka did not
234 successfully reproduce the magnitude and duration of the CDJ+. Additional model
235 experiments would need to be performed to corroborate the influence of obliquity on CDJ+
236 events. To conclude, our study confirms a joint contribution of terrestrial and oceanic carbon
237 release during CDJ- events and points toward the terrestrial carbon reservoir as a variable
238 sensitive to obliquity in this centennial-scale response.

239 Under the conditions of the glacial-interglacial cycles during the last 600 ka, a strong imprint
240 of both obliquity ^{35,36,38,39} and precession ^{40,41} was demonstrated for regional variability of
241 vegetation and climate. Our CDJs compilation reveals no direct correlation between
242 precession and CDJ occurrences over the past 500 ka (Supplementary Figure 5), and our
243 simulations confirm the quantitative impact of obliquity on centennial-scale variability of
244 terrestrial carbon fluxes during CDJ- (Fig. 4). Still, future simulations investigating if changes in
245 precession influence centennial-scale atmospheric CO₂ release would be useful.

246 Our results reinforce the importance of producing carbon isotope and oceanic pH records
247 during CDJ periods to confirm the processes leading to centennial-scale atmospheric CO₂

248 increases. In addition, further modelling studies of both CDJ+ and CDJ- events should be
249 conducted to confirm the pervasive character of the identified mechanism. Here we show that
250 past centennial-scale atmospheric CO₂ increases with associated AMOC disruption were more
251 likely to happen during high-obliquity phases. Five of the past CDJs occurred under an obliquity
252 value that is close to the current one ($\pm 5\%$ of the range of the obliquity value variability of the
253 last 500 ka). While a 15% weakening of the AMOC since the mid-twentieth century has been
254 estimated ⁴², large uncertainties remain on the potential occurrence of a large AMOC
255 disruption at multi-centennial-scale in the future ^{43,44}. If such AMOC disruption occurs, the
256 current relatively high-obliquity phase is favourable to a measurable release of carbon into
257 the atmosphere which would superimpose onto the anthropogenic-sourced perturbation.
258 This release, which could last from a few decades to a few centuries, would be equivalent to
259 ~ 4 years of CO₂ anthropogenic emission at the 2010-2019 emission rate.

260

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286 experimental analyses. LM performed the modelling work. EL and EC wrote a draft with
287 subsequent inputs from all co-authors.

288 **Competing interests:** Authors declare that they have no competing interests

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304 **Figures:**

305 **Fig. 1: Improvement of the temporal resolution of the ice-core atmospheric CO₂ record since the composite**
306 **CO₂ record from ref. ⁴⁵ over the last 500 ka.** a. Top: High-resolution records are represented in blue ¹, red (this
307 study), brown ¹², green ⁸, dark green ⁴⁶, grey ⁴⁷, and light-green dots ⁴⁸, plotted on their own published age scales.
308 They are superimposed on the composite CO₂ record ⁴⁵ (black line). Bottom: Temporal resolutions of all records
309 computed as the running mean of 20 data points (colour codes similar to top panel). Data are not corrected for
310 potential offsets due to the use of different experimental setups. Vertical red bands indicate periods where the
311 ice-core CO₂ record is associated with a temporal resolution better than 500 years. **b.** Top: Zoom on our EDC CO₂
312 record between 260-190 ka (red dots, black line) and CH₄ record from this study (light green) and previous studies
313 ^{49,50} (dark green) plotted on the AICC2023 timescale ²². Vertical black dashed lines indicate the newly identified
314 CDJs (see methods). Middle: CO₂ rates of change from the detrended CO₂ record for the smoothing spline (1 kyr
315 cutoff period) ¹. Bottom: Residual CO₂ amplitude of the detrended CO₂ record during the events occurring at a
316 rate higher than 1.5ppm/century. Vertical back dashed lines indicate the timing of the identified CDJs. A
317 centennial-scale CO₂ release is identified when the rate is higher than 1.5ppm/century and the amplitude is
318 higher than 5 ppm (dashed horizontal red lines) ¹. MIS: Marine Isotope Stage.

319

320 **Fig. 2: Climatic and obliquity states for centennial-scale releases of atmospheric CO₂ during the three periods**
321 **associated with high-resolution ice-core CO₂ records.** a. EDC Deuterium record (δD) ⁵¹ on the AICC2023 ice age
322 scale ²² (blue line). b. Atmospheric CO₂ high-resolution records on the AICC2023 gas age scale ²² for the time
323 interval 500-67 ka, on the WD2014 ^{52,53} time scale for the 67-12 ka interval, and on several original timescales for
324 the 12-0 ka interval (this study; refs. ^{1,8,12,45-48}, red line). c. Obliquity ²⁴ (black). Horizontal dashed lines indicate
325 the average value for each considered proxy record over the last 500 ka. Yellow, orange and white dots indicate
326 the value of the record during the occurrence of CDJ-, CDJ+ and undetermined CDJ, respectively ^{1,8,17}. The CDJ-
327 identified during HS 5 (~49 ka) and simulated with LOVECLIM is identified with black stars on the three
328 represented climate variables.

329

330 **Fig. 3 Simulated carbon cycle response to an AMOC shutdown.** a. Evolution of atmospheric CO₂ for the eight
331 simulations as a function of time. The dashed grey line corresponds to a 1.5 ppm/century CO₂ increase, which is
332 also the threshold used to identify CDJ events in the ice-core CO₂ record. Bold lines are smoothing spline filters.
333 **b.** ΔCO_2 between *HighObl* – *LowObl* (black line), *HighObl_SHW* - *LowObl_SHW* (red line), *HighObl_NoVeg* –
334 *LowObl_NoVeg* (green line) and *HighObl_NoVeg_SHW* – *LowObl_NoVeg_SHW* (blue line). A positive value means
335 that the CO₂ anomaly due to the perturbation is higher under a high-obliquity phase than under a low-obliquity
336 phase. Grey dots with dashed line correspond to the atmospheric CO₂ record from the WAIS ice core during HS
337 5, the HS that is simulated in the present study. **c.** Paired coloured bars correspond to the atmospheric CO₂ offset
338 values 400 years after the freshwater perturbation for simulations using the obliquity value at 49 ka (left, 24.3°)
339 and a prescribed low-obliquity value (right, 22.1°). Words in italic refer to simulation names. **d.** Same simulation
340 results as in panel c but with carbon stock in GtC. Negative carbon anomaly represents a carbon release from
341 that reservoir into the atmosphere. Plain coloured bars represent the oceanic carbon stock and coloured bars
342 with white dots represent the terrestrial carbon stock. Colors refer to the same simulation as in c. The year 400
343 snapshot represents the period with the largest atmospheric CO₂ difference between the high and low obliquity
344 simulations and corresponds to a time period where most of the CO₂ response to the freshwater forcing has
345 occurred (Supplementary Figure 10).

346

347

348 **Fig. 4. Simulated impacts of the freshwater forcing and the obliquity state on the climatic conditions and**
349 **terrestrial carbon stock.** a. Air temperature anomalies resulting from (top) a freshwater forcing under a high

350 obliquity phase vs. no freshwater forcing under a high obliquity phase (*HighObl – HighOblCTR*), (middle) a low
351 obliquity phase without a freshwater forcing vs. a high obliquity phase without a freshwater forcing (*LowOblCTR*
352 – *HighOblCTR*) and (*bottom*) the impact of the freshwater forcing in a high vs. low obliquity phase (*HighObl -*
353 *HighOblCTR*) – (*LowObl - LowOblCTR*) on air temperatures. b. Same as (a) for precipitation anomalies. c. Same as
354 (a) for the terrestrial carbon anomalies. Basemap outlines are from ETOPO5 Global relief
355 model(<https://www.ncei.noaa.gov/products/etopo-global-relief-model>).

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482 **Methods:**

483 **Atmospheric CO₂ measurements on the EDC ice core**

484 Atmospheric CO₂ measurements on the EDC ice core have been performed at Institut des
485 Geosciences de l'Environnement (IGE) using the ball mill dry extraction system coupled with a
486 gas chromatograph^{12,13}. A total of 203 depth levels have been measured between 2043.8 and
487 2360.6 m corresponding to the time period from 258.6 to 190.0 ka on the AICC2023 gas
488 timescale²² (Fig. 1). Each data point corresponds to air extracted from a single ice sample of
489 40 g and the resulting CO₂ amount fraction is the average of five successive injections of the
490 same air sample made at different pressure levels in the gas chromatograph. The pooled
491 standard deviation of the five injections is 0.9 ppm. Thirty injections of a standard gas
492 (synthetic air from Air Liquide) containing 229.5 ± 0.3 ppm of CO₂ in dry air, which was
493 referenced to two primary standards (241.24 ± 0.015 ppm from NOAA, JB03097, and
494 260.26 ± 0.2 ppm from CSIRO, CSIRO1677), are used to calibrate the gas chromatograph. First,
495 we correct our data for the crushing process CO₂ contamination that is estimated to be on
496 average of 2.6 ± 0.9 ppm (1σ standard deviation), ranging from 1.0 to 5.4 ppm. This correction
497 is necessary as some CO₂ is released from the CO₂ extraction chamber during the crushing
498 process, which is contaminating our extracted air from the ice sample. Blank tests are
499 conducted by measuring the CO₂ concentration of a standard gas of known amount fraction
500 and crushing artificial gas-free ice. The difference between the measured and expected CO₂
501 amount fraction is attributed to the crushing process itself. The additional CO₂ amount
502 fraction is then converted into CO₂ levels to normalize the correction to the volume of injected
503 gas. Blank measurements were performed for every ten measured samples for each of the
504 four extraction cells. Following the procedure of ref.¹², the CO₂ data are then corrected for
505 gravitational fractionation occurring in the firn. For that we use a record of the isotopic
506 composition of nitrogen in the trapped air covering our studied time interval ($\delta^{15}\text{N}$ of N₂)⁵⁴
507 (not shown here). The average correction is -1.4 ± 0.3 ppm (1σ standard deviation), ranging
508 from -0.9 to -2.1 ppm. The average overall propagated uncertainty of the CO₂ amount
509 fractions is 1.3 ppm. This uncertainty includes the uncertainty related to the five injections
510 (one standard deviation), the uncertainty of the measurement system as determined by
511 standard gas over blank ice measurements (one standard deviation), and the uncertainty
512 associated with and the correction for gravitational settling based on $\delta^{15}\text{N}$ corrections.

513 Over the 260-190 ka time interval, the existing composite CO₂ record ⁴⁵ includes data
514 measured on the Vostok ice core, associated with a temporal resolution of ~ 0.85 ka ¹³
515 (Supplementary Figure 1). The comparison between our new EDC and the Vostok CO₂ records
516 reveals an offset already observed in ref. ¹² where these data had been measured using an
517 earlier version of the Ball Mill extraction system. Especially, the Vostok CO₂ concentrations are
518 slightly higher by 5.5 ppm on average. This offset can be explained by the fact that the Vostok
519 CO₂ measurements were not corrected for the gravitational fractionation and no blank
520 correction was applied ¹³. When we apply these two corrections on the Vostok CO₂ record,
521 the observed offset is reduced to ~ 1.5 ppm (Supplementary Figure 1).

522 **High resolution CO₂ data compilation**

523 In addition to this study, four multi-centennial-scale resolved atmospheric CO₂ datasets have
524 been published since the atmospheric CO₂ composite record from ref. ⁴⁵ was compiled (Fig.
525 1). Data presented in Fig. 1 are plotted as published and no corrections have been applied.
526 However, this does not affect the results of our study as we focus our analysis on the relative
527 and not absolute variations of CO₂ amount fractions. If we add the duration of all periods with
528 a resolution better than 0.5 ka, we obtain a cumulative duration of 275 ka (vertical red areas,
529 Fig. 1) covering 429 to 329 ka ¹, 258 to 200 ka (this study), 188 to 153 ka ¹², and 84 to 0 ka ^{8,46–}
530 ^{48,55–57}. These periods associated with sub-millennial-scale CO₂ resolution allow to identify
531 centennial scale abrupt variability of atmospheric CO₂ (Fig. 2). In the case of ref. ¹², results of
532 the CDJ identification should be taken with caution as experimental-induced noise make it
533 hard to distinguish natural centennial-scale variability from measurement-induced noise
534 (Supplementary Figure 11). Especially, potential CDJs during this time interval could have been
535 missed. The point-to-point variability registered during some periods of the record (186-183
536 ka; 170-168 ka) confirms that any interpretation of centennial-scale variability during this
537 period needs to carefully take into account this higher variability. We consider that apart from
538 the 429-329 ka, 258-200 ka, 188 to 153 ka and the 84-0 ka periods (Fig. 2), potential CDJ
539 occurrences are not identifiable due to insufficient temporal resolution and larger uncertainty
540 associated with the Vostok data. In addition, ice older than 500 ka is affected by strong
541 thinning and higher resolution CO₂ measurements should be performed in the future, in
542 particular taking advantage of new semi-continuous analytical techniques ⁵⁸.

543 **Atmospheric CH₄ measurements on the EDC ice core**

544 CDJ+ have been linked to atmospheric CH₄ releases of more than 50 ppb¹ in the ice core
545 record. However, their identification requires multi-centennial scale resolution of the CH₄
546 record. Three of the CDJs identified in this study show a potential CH₄ variability but the
547 associated resolution (> 0.5 kyr) was not good enough to validate the occurrence of a
548 synchronous CH₄ increase. The existing CH₄ EDC record for Marine Isotopic Stage (MIS) 8a –
549 6e is based on measurements performed on the IGE and the University of Bern experimental
550 setups⁴⁹. We perform 18 measurements on the time interval of the CDJ+7a, CDJ 7b and CDJ-
551 8a.2, in order to detect a potential CH₄ rise associated with these events. These CH₄
552 measurements on the EDC ice core were performed at IGE using a melt-refreezing extraction
553 method coupled with a gas chromatograph. The method used for the analysis of these samples
554 is identical to that described in ref.⁴⁹. Each sample represents ~40 g of ice from which the air
555 is extracted with a melt-refreezing method. The air is then analysed with a gas chromatograph.
556 15 injections of a single standard gas (synthetic air from Air Liquide) containing 498.1 ± 2.7
557 ppb of CH₄ in dry air, which was referenced to one primary standard 393.9 ± 2.6 ppb from
558 CSIRO, CSIRO1657), are used to calibrate the gas chromatograph. The average CH₄ analytical
559 uncertainty is 5 ppb (1 sigma). We do not correct the concentration for gravitational
560 enrichment to avoid an offset with the previously published record from ref.⁴⁹. Corrections
561 would be ~3 ppb on average, less than 1% of the CH₄ concentrations signal⁴⁹. A total of 18
562 depth levels has been measured focusing on the depth interval where centennial-scale
563 atmospheric CO₂ releases have been identified. All samples are corrected for blank
564 contributions using gas-free ice. All CH₄ data are shown in Supplementary Figure 1.

565 **Identification of the centennial-scale CO₂ events**

566 In order to identify the presence of CDJs in our CO₂ record between 260 and 190 ka, we follow
567 the method developed by ref.¹. We can summarize this methodology as follows: first, we
568 remove the orbital- and millennial-scale trend in the CO₂ record using smoothing splines¹⁴
569 with a cut-off frequency ranging from 6 to 14 ka to evaluate the best detrending method. The
570 definition of the cut-off frequency is derived from the equation (9) of ref.¹⁴. It corresponds to
571 the period at which 50% of the signal is attenuated by the spline-fitting process. The aim is to
572 find a good compromise between removing multi-millennial- to orbital-scale trends while
573 preserving the centennial scale variability (Supplementary Figure 2). Then, a second series of
574 splines ranging from 0.4 to 1.2 ka is applied on the detrended CO₂ record in order to obtain

575 evenly-spaced records aiming at smoothing out measurement uncertainty while still
576 preserving the centennial scale variability. Third, the rate of change of the final residual CO₂
577 record is computed (Fig. 3, Supplementary Figure 2). To identify a CDJ, we apply the thresholds
578 of 1.5 ppm/century and 5 ppm as in ref. ¹. For the 188 to 153 ka ¹² and 84 to 0 ka ^{8,46–48,55–57}
579 intervals, we use the exact same method of identification (Supplementary Figures 11, 12, 13).
580 For the 429 to 329 ka interval, we compile the CDJs already identified using the same method
581 ¹.

582 Other detection methods of centennial-scale variability have previously been applied during
583 the last glacial period and the last deglaciation. The approach from ref. ¹⁷ is based on the
584 analysis of the raw curve of CO₂ concentrations without any corrections to remove the orbital-
585 or the millennial-scale trends. Such an approach is appropriate to identify the centennial-scale
586 atmospheric CO₂ changes of the last termination as the record is very highly resolved (<0.1
587 ka). In addition, the centennial-scale variability is of such magnitude (~12 ppm) that it prevents
588 any error of identification. Recently, ref. ⁸ applied a new approach for the identification of the
589 centennial-scale atmospheric CO₂ variability, using the δD record measured on the ice. Based
590 on the assumption that this local surface temperature proxy does not include centennial-scale
591 climate variability during the studied intervals ¹⁷, they use the δD as a filter for identifying
592 centennial-scale CO₂ release. The first step is to estimate the phase relationship between CO₂
593 and δD in order to align the two records. Then, by subtracting a normalised δD curve to the
594 CO₂ record, they obtain a residual CO₂ signal. This method is well adapted to identify
595 centennial-scale CO₂ release during the last glacial period, where millennial-scale variability is
596 the dominant climatic pattern. Such an approach assumes that δD is a template for millennial-
597 scale CO₂ variability, which is not the case for all time periods, especially during deglaciations
598 ^{59–61} and the end of interglacials ⁶². The coupling between δD and CO₂ during the 260-190 ka
599 interval, which is mainly composed of a deglaciation and an interglacial period, is thus not
600 trivial. In addition, with that method, all centennial-scale variability of carbon cycle is
601 investigated. Hence, we consider that the method of ref. ¹ based on the smoothing spline is
602 most adapted for the identification of the largest centennial-scale atmospheric CO₂ releases,
603 i.e. the CDJ events.

604 Note that applying the ref. ¹ methodology to the most recent period (70-0 ka) where the
605 methodology of ref. ⁸ was originally used to identify potential CDJs (Supplementary Figure 13),

606 led to the same results. Also, the coherency between the two methodologies used for the
607 identification of centennial-scale CO₂ release underlines that the main result of the study, i.e.
608 the influence of the obliquity on the occurrence of such events, is not dependent on the choice
609 of the CDJ detection method.

610 Centennial-scale variability in a recent atmospheric CO₂ record measured on the Taylor glacier
611 ice core has been observed during the Marine Isotope Stage 5-4 transition⁴⁶. However, the
612 short time interval covered by the high-resolution CO₂ record (~13 ka) and the absence of
613 water isotope measurements in the same ice core prevent to apply directly the CDJ
614 methodology identification from ref.¹ or from ref.⁸. To circumvent this issue, we merged the
615 record from ref.⁴⁶ and refs.^{57,63} to produce a continuous record over the 90-45 ka interval
616 (Supplementary Figure 14). We then applied the method of ref.¹ to this merged record. Note
617 that while applied to the whole interval, this test would only allow to identify CDJs on the 74-
618 59 ka interval, where the atmospheric CO₂ concentration resolution is below 500 years⁴⁶ (Fig.
619 1, Supplementary Figure 12).

620 **Chi-square test**

621 We apply a statistical method to investigate the potential role of the Earth's obliquity in the
622 centennial-scale atmospheric CO₂ variability. The Chi-square method allows to test the null
623 hypothesis that a distribution of a parameter (centennial-scale variability occurrences) versus
624 another (obliquity value) is independent. This method assumes a normal distribution of the
625 probability results for the computation of the confidence level. Such approach was applied to
626 determine the impact of external forcing on the climate system in previous studies⁶⁴⁻⁶⁶. For
627 that, we separate the centennial-scale events in two categories depending on whether they
628 occur in the context of an obliquity value above or under the average obliquity of the 12
629 obliquity cycles of last 500 ka (corresponding to the mean obliquity value of the 494.5-9.5 ka
630 period, i.e. 23.33°).

631 For that, we compute:

$$632 \quad X = \frac{(\textit{observed} - \textit{expected})^2}{\textit{expected}}$$

633 With *observed* the number of centennial-scale events occurring below the average obliquity
634 value of the 12 obliquity cycles of the last 500 ka, here 4, and *expected* the theoretical value

635 expected under the null hypothesis. The theoretical value would be 11 (i.e. half of the total
636 number of detected CDJs) if the studied period would contain as many high obliquity intervals
637 than low obliquity periods. However, only 44.1 % of the studied period is below the average
638 obliquity value taken as reference (23.33°). The theoretical value is thus 9.7. We found:

$$639 \quad X = 3.4$$

640 This value is higher than $\alpha_{0,1}$, the risk at 10% to accept the null hypothesis even if it is wrong,
641 which is 2.7 for a degree of freedom of 1. In other words, the value 2.7 is the 90th percentile
642 of the chi-square distribution ⁶⁷.

643 We can thus reject the null hypothesis that the distribution of centennial-scale CO₂ events is
644 independent of the obliquity state with a confidence level of 90%. In order to test the
645 sensitivity of the chosen threshold (here the average obliquity value of the 12 obliquity cycles
646 of the last 500 ka), we performed a chi-square test for the interval of obliquity values (22.7 -
647 23.9°) systematically reached during each obliquity cycle over the last 500 ka (Supplementary
648 Figure 16). Our analysis reinforces the observation of missing CDJ during low-obliquity periods.

649 We also apply the Chi-square test for the Earth's precession and eccentricity parameters using
650 the mean values of the last 500 ka (Supplementary Table 2). The null hypothesis of an
651 independence of centennial-scale CO₂ events from precession and eccentricity context could
652 not be rejected at a confidence level of 90%.

653 **Experimental design of carbon cycle modelling**

654 To understand the processes that could lead to a different response of the carbon cycle during
655 an AMOC shutdown as a function of obliquity, numerical experiments are performed with the
656 Earth system model LOVECLIM²⁶ (Supplementary Table 4). LOVECLIM includes a free surface
657 primitive equation ocean model with a horizontal resolution of 3°x3° and 20 unevenly spaced
658 vertical levels, a dynamic/thermodynamic sea-ice model, a quasi-geostrophic T21
659 atmospheric model, a dynamic global vegetation model (DGVM) and a marine carbon cycle
660 model. The DGVM used, VECODE (Vegetation Continuous Description model), simulates the
661 vegetation structure, and associated terrestrial carbon reservoirs ²⁶.

662 The initial 49 ka state was derived from a transient experiment that started at 140 ka. The
663 penultimate deglaciation (140-120 ka) followed the PMIP4 protocol ⁶⁸, with the model being

664 forced by changes in orbital parameters ⁶⁹, atmospheric greenhouse gases ⁷⁰ and changes in
665 continental ice-sheets ⁶⁸. That simulation was continued between 120 and 49 ka forced with
666 estimates of changes in continental ice-sheets from an offline simulation ⁷¹. The Bering strait
667 was gradually closed at 70 ka, and kept closed until the early Holocene.

668 From that initial state at 49 ka, the model was run for 4000 years with prognostic atmospheric
669 CO₂ concentration and under constant 49 ka boundary conditions (HighOblCTR). The last 1000
670 yrs of this experiment serve as control run for the 49 ka experiments (Supplementary Figure
671 17).

672 From the initial transient state, another experiment was run under 49 ka boundary conditions
673 but with an obliquity of 22.1°. This experiment was first run for 4000 years with a constant
674 atmospheric CO₂ concentration of 204 ppm, after which the CO₂ was prognostic for 2000 years
675 (LowOblCTR) (Supplementary Figure 17).

676 From these control runs (HighOblCTR and LowOblCTR), two meltwater experiments were
677 performed (HighObl and LowObl) by adding 0.3 Sv of meltwater into the North Atlantic (60-
678 10°W, 50-60°N) for 1000 years thus leading to an AMOC shutdown in ~300 years. The
679 meltwater flux was then stopped for 800 years, after which a negative meltwater flux was
680 added to the North Atlantic (-0.1 Sv) for 400 years so that the AMOC would recover.

681 Two experiments with similar forcing to HighObl and LowObl but with a 40% increase of the
682 westerly wind stress over the Southern Ocean between 65°S and 34°S from year 0 to 1800
683 (HighObl_SHW and LowObl_SHW) were also performed.

684 Finally, two experiments similar to HighObl and LowObl, and two experiments similar to
685 HighObl_SHW and LowObl_SHW were performed, in which the terrestrial carbon fluxes did
686 not impact atmospheric CO₂ (HighObl_NoVeg, LowObl_NoVeg, HighObl_NoVeg_SHW and
687 LowObl_NoVeg_SHW). Note that for these experiments, the model was first run for 2000
688 years without any meltwater so that the atmospheric CO₂ concentration would equilibrate.

689 We also perform some numerical experiments to better understand the carbon cycle response
690 to CDJ+. This is however challenging as the climatic conditions associated with CDJ+ are poorly
691 known, implying that the background climatic conditions and AMOC change associated with
692 these events are poorly constrained. Nevertheless, one CDJ+ is identified at the end of the
693 Younger Dryas, when the obliquity is high (24.16°). Starting from a transient simulation of the

694 last deglaciation³², we branch an equilibrium simulation forced with constant boundary
695 conditions (i.e. orbital parameters, ice-sheet configuration and albedo, Bering Strait open)
696 representing 12 ka but with interactive atmospheric CO₂ (YDctr). A similar equilibrium
697 experiment is performed but with low obliquity (22.1°, YD-lowOblctr). From these equilibrium
698 states, 0.16 Sv of freshwater is added into the North Atlantic for 1000 years to simulate an
699 AMOC weakening (YDNA and YDlowObINA) (Supplementary Figure 21).

700

701 **Data availability:** The atmospheric CO₂ and CH₄ data are available at

702 <https://doi.pangaea.de/10.1594/PANGAEA.972146> (ref. 70) and

703 <https://doi.pangaea.de/10.1594/PANGAEA.972147> (ref. 71)

704 The output simulations are available at <http://hdl.handle.net/1959.4/102810> (DOI:

705 <https://doi.org/10.26190/unsworks/30408>) (ref. 72)

706 **Code availability:** The simulations are based on the LOVECLIM model version 1.2, whose

707 code is accessible at <http://www.elic.ucl.ac.be/modx/elic/index.php?id=289> (last access: 05

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709

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