



HAL
open science

Carbonate platform production during the Cretaceous.

Alexandre Pohl, Yannick Donnadiou, Yves Godderis, Cyprien Lanteaume,
Alex Hairabian, Camille Frau, Julien Michel, Marie Laugie, John J.G.
Reijmer, Christopher R. Scotese, et al.

► **To cite this version:**

Alexandre Pohl, Yannick Donnadiou, Yves Godderis, Cyprien Lanteaume, Alex Hairabian, et al..
Carbonate platform production during the Cretaceous.. Geological Society of America Bulletin, 2020,
132 (11-12), pp.2606-2610. 10.1130/B35680.1 . hal-03003984

HAL Id: hal-03003984

<https://hal.science/hal-03003984v1>

Submitted on 3 Sep 2021

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Carbonate platform production during the Cretaceous

Alexandre POHL^{1,2,3}, Yannick DONNADIEU¹, Yves GODDERIS⁴, Cyprien LANTEAUME^{1,5,6}, Alex HAIRABIAN⁷, Camille FRAU¹, Julien MICHEL^{1,5}, Marie LAUGIE¹, John J. G. REIJMER⁷, Christopher R. SCOTese⁸, Jean BORGOMANO¹.

¹*Aix Marseille Univ, CNRS, IRD, Coll France, INRA, CEREGE. Aix-en-Provence, France*

²*Department of Earth Sciences, University of California, Riverside, CA, USA*

³*Biogéosciences, UMR 6282, UBFC/CNRS, Université Bourgogne Franche-Comté, 6 boulevard Gabriel, F-21000 Dijon, France*

⁴*UMR5563 Géosciences Environnement Toulouse, Observatoire Midi-Pyrénées, CNRS, Toulouse, France*

⁵*MODIS Pau, 4 Rue Jules Ferry, Pau, France*

⁶*Total CSTJF, Avenue Larribeau, 64000 Pau, France*

⁷*College of Petroleum Engineering & Geosciences, King Fahd University of Petroleum & Minerals, Dhahran 31261, Saudi Arabia*

⁸*Department of Earth & Planetary Sciences, Northwestern University, Evanston, IL, 60201, USA*

ABSTRACT

Platform carbonates are among the most voluminous of Cretaceous deposits. The production of carbonate platforms fluctuated through time. Yet, the reasons for these fluctuations are not well understood and the underlying mechanisms remain largely unconstrained. Here we document the long-term trend in Cretaceous carbonate platform preservation based on a new data compilation and use a climate-carbon cycle model to explore the drivers of carbonate platform production during the Cretaceous. We show that neritic carbonate preservation rates followed a

24 unimodal pattern during the Cretaceous reaching maximum values during the mid Cretaceous
25 (Albian, 110 Ma). Coupled climate-carbon cycle modeling reveals that this maximum in
26 carbonate deposition results from a unique combination of high volcanic degassing rates and
27 widespread shallow-marine environments that served as a substrate for neritic carbonate
28 deposition. Our experiments demonstrate that the unimodal pattern in neritic carbonate
29 accumulation agrees well with most of the volcanic degassing scenarios for the Cretaceous. Our
30 results suggest that the first-order temporal evolution of neritic carbonate production during the
31 Cretaceous reflects changes in continental configuration and volcanic degassing. Geodynamics,
32 by modulating accommodation space, and turnovers in the dominant biota probably played a role
33 as well, but it is not necessary to account for the latter processes to explain the first-order trend in
34 Cretaceous neritic carbonate accumulation in our simulations.

35 **INTRODUCTION**

36 In the recent years, our knowledge of the geological evolution of the surficial carbon cycle
37 has greatly improved. Among the main advances is the reconstruction of the Phanerozoic
38 atmospheric CO₂ (Bernier, 2006; Foster et al., 2017). The fluctuations of the carbon sink due to
39 continental weathering as well as the change in solid Earth degassing have been at least partly
40 constrained, using models (Goddéris et al., 2014; Brune et al., 2017) or proxies (such as isotopes)
41 (Veizer et al., 1999; Cao et al., 2017). One major component of the carbon cycle, however, the
42 sedimentary carbonate sink, has been poorly investigated at the global scale. Many data have
43 been published, regarding carbonate accumulation fluxes (Bosscher and Schlager, 1993;
44 Kiessling et al., 2000), but data compilations and extrapolation at the global scale are scarce.

45 Carbonate sediments are the receptacle of the atmospheric CO₂ that has been removed
46 from the atmosphere by weathering, which can be subsequently recycled along subduction zones

47 (Mason et al., 2017; Pall et al., 2018). Carbonate sediment deposition thus reflects many coupled
48 processes. Identifying those processes through geological times is an unparalleled way to check
49 whether we have understood the overall functioning of the carbon cycle or not.

50 Here we investigate the production by carbonate platforms during the Cretaceous, one of
51 the most prolific periods of carbonate platform development (Kiessling et al., 2003). We
52 reconstruct temporal variations in platform carbonate deposition based on a new data
53 compilation and subsequently explore the drivers of the temporal trends in Cretaceous carbonate
54 production using a coupled climate-carbon cycle model.

55 **METHODS**

56 **Database of shallow-water carbonate preservation rates**

57 We compiled, from published studies, an extensive database of Cretaceous shallow-water
58 marine carbonate preservation rates (see the GSA Data Repository for the full database). The
59 latter were calculated by dividing the estimated sediment thickness by the stratigraphic interval
60 duration. Calculated preservation rates necessarily underestimate actual carbonate production
61 rates. Indeed, neritic carbonate accumulation is limited by depositional accommodation space. In
62 addition, deposits may subsequently undergo dissolution, erosion and compaction, and produced
63 sediments may be exported toward the basin. As a result of all those mechanisms acting to
64 reduce the thickness of the carbonates preserved on the shelves, preservation rates calculated
65 based on the geological record are typically one to several orders of magnitude lower than
66 production rates measured on present-day carbonate-secreting ecosystems such as coral reefs
67 (Bosscher, 1992; Strasser and Samankassou, 2003; Schlager, 2010). Therefore compiled
68 preservation rates are considered significant underestimations of the actual accumulation rates.
69 Considering the multiple mechanisms acting to reduce the thickness of the shallow-water

70 carbonates preserved on the shelves, maximum preservation rates are further considered to be the
71 most representative of the original carbonate accumulation flux in ancient times.

72 **The GEOCLIM climate-carbon cycle model**

73 We used the GEOCLIM climate-carbon cycle model (see the Data Repository) to simulate
74 carbonate accumulation fluxes during the Cretaceous. The GEOCLIM model (Donnadieu et al.,
75 2006a) asynchronously couples a carbon cycle box model with a general circulation model. It
76 represents a significant improvement over the zero-dimensional GEOCARB-like models (Berner,
77 2006), since continental weathering rates are resolved spatially, which takes into account the
78 continental configuration and sea level (Goddéris et al., 2014). The carbon cycle box model is the
79 COMBINE model upgraded following Donnadieu et al. (2006a). It includes a robust
80 representation of the marine sub-cycles of inorganic and organic carbon, oxygen, alkalinity and
81 phosphorus. It explicitly represents the spatial distribution of silicate, basalt and carbonate
82 weathering as a function of climate. The oceans are modeled by 9 “boxes”. The accumulation of
83 neritic carbonates takes place exclusively in the box representing the low-to-mid latitude
84 epicontinental photic zone, following Eqn. 1:

$$85 \quad F = k_{cr} \times A_{platform} \times (\Omega_{ara} - 1)^{1.7} \quad (1)$$

86 where F is the neritic carbonate depositional flux (mol yr^{-1}), $A_{platform}$ is the shelf area
87 available to shallow-water carbonate deposition, derived from the paleogeographic
88 reconstructions, Ω_{ara} is the model aragonite solubility ratio and k_{cr} a calibration constant. This
89 formulation of neritic carbonate accumulation rates has been successfully used in previous
90 studies to investigate the consequences of the demise of carbonate platforms during the Middle–
91 Late Jurassic Transition (ca. 160 Ma; Donnadieu et al., 2011) and the effects of the breakup of
92 Pangea on neritic carbonate accumulation (Late Permian to Late Cretaceous; Donnadieu et al.,

93 2006a). The climatic component of GEOCLIM is the FOAM ocean-atmosphere model version
94 1.5 (Jacob, 1997), a mixed-resolution general circulation model that has been routinely used to
95 study climate in deep time, including during the Cretaceous (Donnadieu et al., 2006b; Ladant and
96 Donnadieu, 2016).

97 Simulations were conducted using the Cretaceous paleogeographical reconstructions of
98 Scotese (2016) and Scotese and Wright (2018) with deep-ocean bathymetry after Müller et al.
99 (2008). Changing solar luminosity values were modeled after the stellar physics of Gough
100 (1981). Earth's orbit around the Sun was circular (null eccentricity) and the obliquity was set to
101 22°. This orbital configuration provides an equal, annual insolation for both hemispheres with
102 minimal seasonal contrast.

103 **RESULTS**

104 **Temporal variations in Cretaceous shallow-water carbonate preservation rates**

105 At first order, Fig. 1 shows a unimodal pattern in neritic carbonate preservation rates
106 during the Cretaceous with background values $< 70 \text{ m Myr}^{-1}$ and an excursion to higher
107 preservation rates up to $175\text{--}200 \text{ m Myr}^{-1}$ during the mid Cretaceous between 125 Ma (Aptian)
108 and 95 Ma (Cenomanian). This mid-Cretaceous maximum is a robust feature. It was already
109 highlighted by the previous Phanerozoic compilation of Bosscher and Schlager (1993). It
110 parallels the mid-Cretaceous local maximum in reefal carbonate production identified by
111 Kiessling et al. (2000). It also corresponds with the development of massive successions of
112 sediments in the Mediterranean Tethys ($>100 \text{ m}$ thick; Skelton and Gili, 2011) and with the
113 maximum in carbonate accumulation identified in the North American, Russian and Siberian
114 platforms (Whitten, 1976).

115 **Coupled climate-carbon cycle simulations**

116 The drivers of secular change in platform carbonate production have been the subject of
117 investigation for decades (Schlager, 1981; Bosscher and Schlager, 1993; Kiessling et al., 2000;
118 Donnadieu et al., 2006a). From a carbon cycle perspective, global carbonate deposition is tightly
119 linked to changes in continental silicate and carbonate weathering. However, several factors
120 impact the distribution of carbonate deposition between the open ocean and shallow-marine
121 platforms. Debate thus continues as to whether fluctuations in neritic carbonate accumulation
122 rates reflect changing environmental conditions such as global climate, evolving geodynamics
123 (e.g., continental configuration and subsidence) or intrinsic changes such as turnovers in
124 dominant biota, since different carbonate factories are known to be associated with different
125 carbonate production rates (e.g., Lanteaume et al., 2018). Previous attempts to identify these
126 complex relationships over the course of the Phanerozoic failed to reach any definitive
127 conclusions (Bosscher and Schlager, 1993; Kiessling et al., 2000).

128 *Baseline simulations with present-day volcanism*

129 We used the GEOCLIM climate-carbon cycle model (Donnadieu et al., 2006a) to
130 investigate the driving mechanisms for the Cretaceous trend in neritic carbonate deposition with
131 a special focus on the maximum in carbonate accumulation centered at 110 Ma (see Fig. 1).
132 Simulations were run every 10 Myrs between 150 Ma (Late Jurassic) and 60 Ma (Paleocene).
133 For each time slice and associated continental configuration, the GEOCLIM model computed the
134 $p\text{CO}_2$ value by balancing volcanic degassing with silicate weathering. The GEOCLIM model
135 simulated the associated climatic state and ocean biogeochemistry. Though the neritic carbonate
136 deposition rate was focus of our investigation, the simulated $p\text{CO}_2$ was used to evaluate the fit of
137 our simulations to the latest $p\text{CO}_2$ reconstruction of Foster et al. (2017). The latter criterion was

138 chosen to compare our simulations to proxy data due to the critical role that $p\text{CO}_2$ plays on ocean
139 pH and thus carbonate geochemistry.

140 In our baseline runs, the rate of volcanic degassing was set to the present-day value ($6.8 \times$
141 10^{12} moles of carbon per year; Donnadieu et al., 2006a) (Fig. 2A); only the paleogeography and
142 associated climate varied from one time slice to another. Simulated $p\text{CO}_2$ was lower than
143 indicated by the proxy compilation for most of the Cretaceous (Fig. 2B). Yet, an interesting
144 pattern emerged: the bimodal trend observed in the $p\text{CO}_2$ compilation between 120 Ma and 60
145 Ma (grey shading in Fig. 2B) was well reproduced in the baseline runs (black line in Fig. 2B).
146 This suggests that the temporal variations in $p\text{CO}_2$ reconstructed by Foster et al. (2017) between
147 120 Ma and 60 Ma may be largely due to the paleogeographical evolution. The total neritic
148 carbonate depositional flux simulated in the baseline runs varied very little between 150 Ma and
149 60 Ma, showing only a local optimum at 80 Ma (Fig. 2C).

150 *Numerical simulations using Cretaceous volcanic degassing scenarios*

151 In a second step, we used Cretaceous plate models and data compilations to derive
152 volcanic CO_2 degassing scenarios, which could help reconcile the simulations with proxy-based
153 $p\text{CO}_2$ reconstructions. Such modeling attempt faces limitations associated with the large
154 uncertainties in the reconstruction of Cretaceous volcanic degassing. Indeed, numerous studies
155 have been published trying to reconstruct one of the components of the degassing rate but there
156 is no reconstruction of the total CO_2 degassing by the solid Earth at the geological timescale. The
157 lack of an integrated curve leaves us with the necessity to rely on incomplete reconstructions.
158 Such bias concerns both the modeling and data communities. For instance, McKenzie et al.
159 (2016) argue for a dominant role of degassing on the climatic evolution, based on a correlation

160 between the Phanerozoic climate and their curve for the degassing of arc volcanism, although arc
161 volcanism is only one component of the total degassing.

162 The rationale that we adopted thus consists in using available constraints to compute total
163 volcanic degassing rates, expressed as ratios compared to the present-day value. We did not try
164 to identify the individual contributions of the different components: subduction zones, rifts, mid-
165 ocean ridges and arc volcanism. Instead, we followed the approach adopted by Berner in his
166 various versions of the seminal GEOCARB(SULF) model (Berner and Kothavala, 2001; Berner,
167 2004, 2006), which consists in scaling volcanic degassing with proxies for geodynamic activity
168 (e.g., subduction length). Although this approach will undoubtedly benefit from advances in
169 geodynamic models in the near future, it has been shown to produce very good results in the
170 GEOCARB model (Berner, 2006; Van Der Meer et al., 2014). In detail, we derived volcanic
171 degassing rates (Fig. 2A) from the length of subduction zones calculated by Van Der Meer et al.
172 (2014) (hereafter VDM14), from the area of subducted lithosphere reconstructed over the last
173 180 Myrs by Engebretson et al. (1992) (E92), from the slab flux estimated by East et al. (2020)
174 (EA20) and from the continental arc activity compiled by Cao et al. (2017) (CAO17). We also
175 used the C1' and C1'' scenarios of Brune et al. (2017), which consider variable contributions of
176 rifts, mid-ocean ridges, subduction zones and continental arcs.

177 Our model does not account for the degassing from large igneous provinces (LIPs). It has
178 been shown (Dessert et al., 2001) that LIP degassing increases the atmospheric partial pressure
179 of CO₂ for durations that do not exceed a few million years after the end of the onset. LIPs also
180 impact the carbonate system through short-term acidification of the ocean, just prior to the
181 response of the weathering system (Paris et al., 2016). However, we here explore the response of
182 the carbon cycle to processes acting at a longer timescale, in the order of ten million years.

183 We tested each of the volcanic degassing scenarios in the GEOCLIM model. The best fit
184 to proxy data was obtained using the subduction-based degassing scenario E92 (Fig. 2B). The
185 neritic carbonate production flux associated with this best-guess scenario displayed a maximum
186 centered on 110 Ma (Fig. 2C) that agreed well with our compilation of Cretaceous preservation
187 rates (see Fig. 1).

188 *Analyzing the drivers of the Mid-Cretaceous maximum in carbonate accumulation*

189 Despite the disparities in the volcanic degassing scenarios (Fig. 2A), most simulations
190 showed that the highest rates of neritic carbonate production occurred during the mid Cretaceous
191 (between ca. 120 and 80 Ma). The mid-Cretaceous unimodal pattern, bounded by an increase in
192 neritic carbonate production in the Early Cretaceous (120-130 Ma) and a decrease in neritic
193 carbonate production in the early Late Cretaceous (80 Ma), was a robust pattern that was shared
194 by different degassing scenarios.

195 Comparison of the simulations that considered evolving volcanic degassing rates with the
196 baseline runs (color vs. black lines in Fig. 2C) revealed that the lowest-to-mid Cretaceous
197 increase in neritic carbonate production (130–120 Ma), at least partially, relied on an increase in
198 volcanic degassing rates. In addition, a sensitivity test that used a constant area of the neritic
199 environments available for carbonate deposition (dashed magenta line in Fig. 2C) demonstrated
200 that the latter area played a key role in the increase in shallow-marine carbonate production.
201 Indeed, the area of the continental shelves significantly rose between 130 Ma and 120 Ma in our
202 reconstructions (Fig. 3). This increase resulted from the strong eustatic rise that took place at that
203 time (Müller et al., 2008). In the E92 scenario, the increasing area of shallow-marine
204 environments strengthened the pattern of carbonate accumulation already imposed by changing
205 volcanism (Fig. 2C), by favoring the accumulation of carbonate in neritic environments at the

206 expense of the open ocean. In other degassing scenarios that did not have a well-defined
207 maximum at 120–110 Ma (i.e., C1', C1'', VDM14, CAO17, EA20), the increasing area of the
208 shallow-marine environments, alone, drove the increase in neritic carbonate production between
209 130 and 120 Ma (Figs. 2A,C). Regarding the decrease in carbonate production after 80 Ma, two
210 main mechanisms were in play; most scenarios showed a drop in volcanic degassing (Fig. 2A)
211 and a major eustatic fall (Müller et al., 2008) reduced the area of the shallow-marine
212 environments available to neritic carbonate deposition (Fig. 3).

213 **DISCUSSION**

214 Comparison of our simulated trend of shallow-water carbonate production with the
215 updated compilation of preservation rates allowed us to propose a mechanistic explanation to
216 temporal variations in the deposition of Cretaceous shallow-water carbonates. It would be even
217 more interesting and straightforward to compare our simulated fluxes with the volumes (instead
218 of vertical aggradation rates) of shallow-water carbonates produced during the Cretaceous.
219 Unfortunately, such comparison is hampered by the lack of robust quantification of these
220 volumes. For instance, Dutkiewicz et al. (2019) calculated platform carbonate accumulation
221 fluxes by multiplying the extent of the carbonate platforms mapped by Kiessling et al. (2003) by
222 globally-uniform preservation rates derived from the compilation of Bosscher and Schlager
223 (1993). Figure 1C reveals that the temporal trends of carbonate production simulated in
224 GEOCLIM and the preserved volumes reconstructed by Dutkiewicz et al. (2019) largely differ.
225 The estimates of Dutkiewicz et al. (2019) provide significantly lower values. The low values at
226 least partly reflect the action of dissolution, erosion, compaction and the export of the carbonates
227 toward the basin, which all tend to reduce the thickness of the carbonates deposited on the
228 shelves. Preserved volumes necessarily represent an underestimation of the volumes of carbonate

229 actually produced (see Methods). Regarding the temporal trend, the reconstruction of Dutkiewicz
230 et al. (2019) does not provide any robust quantification of the variation in the production of
231 platform carbonates. Indeed, the approach of Dutkiewicz et al. (2019) consists in multiplying the
232 surface of the carbonate platforms documented in the geological database by a constant (i.e., a
233 uniform rate of vertical aggradation). The authors derived two values for this constant rate from
234 the compilation of Bosscher and Schlager (1993), which provided the lower (using 30 m Myr⁻¹)
235 and upper (using 80 m Myr⁻¹) boundaries for their reconstructed trend (Fig. 1C). Our updated
236 compilation of preservation rates (Fig. 1), however, reveals that preservation rates cover a very
237 large range of values, from < 10 to 200 m Myr⁻¹. While it probably reflects the best estimate that
238 can be produced at the global scale based on the current knowledge, the approach employed by
239 Dutkiewicz et al. (2019) neglects the spatio-temporal variations in carbonate accumulation rates
240 and thus exclusively reflects the evolution of the area of the carbonate platforms (compare Fig.
241 1C and Fig. 3). Skelton (2003) conducted similar calculations for the Aptian and Cenomanian.
242 Only the platforms of the Tethyan region were considered in these calculations, while
243 Dutkiewicz et al. (2019) considered the area of the platforms at the global scale. The fact that
244 both studies led to virtually identical values (Fig. 1C), although they focused on significantly
245 different spatial scales, highlights once again the difficulty to estimate the depositional flux by
246 neritic carbonates in the deep time. We argue that, considering the current difficulty to estimate
247 volumes of neritic carbonates produced in ancient times, our reconstructed neritic carbonate
248 preservation rates (Fig. 1) constitute an instructive proxy for the efficiency of the shallow-water
249 carbonate factory in ancient times. The reported temporal patterns are supported by both
250 previous data compilations (Whitten, 1976; Bosscher and Schlager, 1993; Kiessling et al., 2000)
251 and accumulation patterns observed in the field (Skelton and Gili, 2011).

252 CONCLUSION

253 The simulated maximum in mid-Cretaceous neritic carbonate production results from the
254 combination of high volcanic degassing levels and the large shallow-marine realms. Although
255 the precise evolution of carbonate accumulation rates varies from one volcanic degassing
256 scenario to another, the unimodal pattern in neritic carbonate production simulated during the
257 Cretaceous is robust and shows up in most of the scenarios. The model results align well with the
258 temporal trend that emerges from the updated compilation of Cretaceous platform carbonate
259 preservation rates. Our results therefore suggest that the first-order evolution of neritic carbonate
260 deposition during the Cretaceous reflects changes in continental configuration and volcanic
261 degassing. Other mechanisms modulating accommodation space and turnovers in the dominant
262 biota probably played a role as well, but our simulations suggest that the latter processes do not
263 explain the first-order trend in Cretaceous neritic carbonate accumulation.

264 ACKNOWLEDGMENTS



265
266 The authors thank Dietmar Müller, Peter Skelton and two anonymous reviewers for their
267 thorough reviews that significantly strengthened the present manuscript, and Judith Totman
268 Parrish and Gerald Dickens for editorial handling. A.P. and Y.D. thank the CEA/CCRT for
269 providing access to the HPC resources of TGCC under the allocation 2014–012212 made by
270 GENCI. The research leading to these results was supported by a Marie Skłodowska-Curie
271 International Fellowship under Grand Agreement no. 838373. The authors thank B. Suchéras-
272 Marx (CEREGE) for interesting discussions on the calcifying plankton during the Cretaceous, P.
273 Maffre (GET) for his assistance in setting-up our Cretaceous configuration of the GEOCLIM
274 model and S. Brune for providing degassing scenarios.

275

276 **REFERENCES CITED**

277 Berner, R.A., 2006, GEOCARBSULF: A combined model for Phanerozoic atmospheric O₂ and
278 CO₂: *Geochimica et Cosmochimica Acta*, v. 70, p. 5653–5664,

279 <http://www.sciencedirect.com/science/article/pii/S0016703706002031>.

280 Berner, R.A., 2004, *The phanerozoic carbon cycle CO₂ s and O₂*: New York, Oxford University
281 Press.

282 Berner, R.A., and Kothavala, Z., 2001, GEOCARB III: a revised model of atmospheric CO₂
283 over Phanerozoic time: *American Journal of Science*, v. 301, p. 182–204,

284 <http://ajsonline.org/content/301/2/182.short>.

285 Bosscher, H., 1992, *Growth potential of coral reefs and carbonate platforms*: 157 p. p.

286 Bosscher, H., and Schlager, W., 1993, Accumulation rates of carbonate platforms: *Journal of*
287 *Geology*, v. 101, p. 345–355, doi:10.1086/648228.

288 Brune, S., Williams, S.E., and Müller, R.D., 2017, Potential links between continental rifting,
289 CO₂ degassing and climate change through time: *Nature Geoscience*, v. 10, p. 941–946,
290 doi:10.1038/s41561-017-0003-6.

291 Cao, W., Lee, C.-T.A., and Lackey, J.S., 2017, Episodic nature of continental arc activity since
292 750 Ma: A global compilation: *Earth and Planetary Science Letters*, v. 461, p. 85–95.

293 Dessert, C., Dupré, B., François, L.M., Schott, J., Gaillardet, J., Chakrapani, G., and Bajpai, S.,
294 2001, Erosion of Deccan Traps determined by river geochemistry: impact on the global
295 climate and the ⁸⁷Sr/⁸⁶Sr ratio of seawater: *Earth and Planetary Science Letters*, v. 188, p.
296 459–474.

297 Donnadiou, Y., Dromart, G., Goddérís, Y., Pucéat, E., Brigaud, B., Dera, G., Dumas, C., and

298 Olivier, N., 2011, A mechanism for brief glacial episodes in the Mesozoic greenhouse:
299 Paleoclimatology, v. 26, p. PA3212, <http://doi.wiley.com/10.1029/2010PA002100>.

300 Donnadieu, Y., Godd ris, Y., Pierrehumbert, R., Dromart, G., Fluteau, F., and Jacob, R., 2006a,
301 A GEOCLIM simulation of climatic and biogeochemical consequences of Pangea breakup:
302 Geochemistry, Geophysics, Geosystems, v. 7, p. Q11019,
303 <http://doi.wiley.com/10.1029/2006GC001278>.

304 Donnadieu, Y., Pierrehumbert, R., Jacob, R., and Fluteau, F., 2006b, Modelling the primary
305 control of paleogeography on Cretaceous climate: Earth and Planetary Science Letters, v.
306 248, p. 426–437.

307 Dutkiewicz, A., M ller, M.D., Cannon, J., Vaughan, S., and Zahirovic, S., 2019, Sequestration
308 and subduction of deep-sea carbonate in the global ocean since the Early Cretaceous:
309 Geology, v. 47, p. 91–94, doi:10.1130/G45424.1.

310 East, M., M ller, R.D., Williams, S., Zahirovic, S., and Heine, C., 2020, Subduction history
311 reveals Cretaceous slab superflux as a possible cause for the mid-Cretaceous plume pulse
312 and superswell events: Gondwana Research, v. 79, p. 125–139,
313 doi:10.1016/j.gr.2019.09.001.

314 Engebretson, D.C., Kelley, K.P., Cashman, H.J., and Richards, M.A., 1992, 180 Million Years of
315 Subduction: GSA Today, v. 2, p. 93–100.

316 Foster, G.L., Royer, D.L., and Lunt, D.J., 2017, Future climate forcing potentially without
317 precedent in the last 420 million years: Nature Communications, v. 8, p. 14845,
318 doi:10.1038/ncomms14845.

319 Godd ris, Y., Donnadieu, Y., Le Hir, G., Lefebvre, V., and Nardin, E., 2014, The role of
320 palaeogeography in the Phanerozoic history of atmospheric CO₂ and climate: Earth-Science

321 Reviews, v. 128, p. 122–138, doi:10.1016/j.earscirev.2013.11.004.

322 Gough, D.O., 1981, Solar interior structure and luminosity variations*: Solar Physics, v. 74, p.

323 21–34.

324 Gradstein, F.M., Ogg, J.G., Schmitz, M., and Ogg, G. (Eds.), 2012, The Geologic Time Scale,

325 Volumes 1 & 2: Elsevier.

326 Jacob, R.L., 1997, Low frequency variability in a simulated atmosphere-ocean system:

327 University of Wisconsin Madison, 170 p.

328 Kiessling, W., Flügel, E., and Golonka, J., 2000, Fluctuations in the carbonate production of

329 Phanerozoic reefs: Geological Society Special Publication, v. 178, p. 191–215,

330 doi:10.1144/GSL.SP.2000.178.01.13.

331 Kiessling, W., Flügel, E., and Golonka, J., 2003, Patterns of phanerozoic carbonate platform

332 sedimentation: v. 36, 195–226 p., doi:10.1080/00241160310004648.

333 Ladant, J.-B., and Donnadieu, Y., 2016, Palaeogeographic regulation of glacial events during the

334 Cretaceous supergreenhouse: Nature communications, v. 7, p. 12771,

335 <http://dx.doi.org/10.1038/ncomms12771>.

336 Lanteaume, C., Fournier, F., Pellerin, M., and Borgomano, J., 2018, Testing geologic

337 assumptions and scenarios in carbonate exploration: Insights from integrated stratigraphic,

338 diagenetic, and seismic forward modeling: Leading Edge, v. 37, p. 672–680,

339 doi:10.1190/tle37090672.1.

340 Mason, E., Edmonds, M., and Turchyn, A. V, 2017, Remobilization of crustal carbon may

341 dominate volcanic arc emissions: Science, v. 294, p. 290–294.

342 McKenzie, N.R., Horton, B.K., Loomis, S.E., Stockli, D.F., Planavsky, N.J., and Lee, C.-T.A.,

343 2016, Continental arc volcanism as the principal driver of icehouse-greenhouse variability:

344 Science, v. 352, p. 444–447.

345 Van Der Meer, D.G., Zeebe, R.E., Van Hinsbergen, D.J.J., Sluijs, A., Spakman, W., and Torsvik,
346 T.H., 2014, Plate tectonic controls on atmospheric CO₂ levels since the Triassic:
347 Proceedings of the National Academy of Sciences of the United States of America, v. 111,
348 p. 4380–4385, doi:10.1073/pnas.1315657111.

349 Müller, R.D., Sdrolias, M., Gaina, C., Steinberger, B., and Heine, C., 2008, Long-term sea-level
350 fluctuations driven by ocean basin dynamics: Science, v. 319, p. 1357–1362,
351 <http://www.sciencemag.org/cgi/doi/10.1126/science.1151540>.

352 Pall, J., Zahirovic, S., Doss, S., Hassan, R., Matthews, K.J., Cannon, J., Gurnis, M., Moresi, L.,
353 Lenardic, A., and Dietmar Müller, R., 2018, The influence of carbonate platform
354 interactions with subduction zone volcanism on palaeo-atmospheric CO₂ since the
355 Devonian: Climate of the Past, v. 14, p. 857–870, doi:10.5194/cp-14-857-2018.

356 Paris, G., Donnadiou, Y., Beaumont, V., Fluteau, F., and Godd ris, Y., 2016, Geochemical
357 consequences of intense pulse-like degassing during the onset of the Central Atlantic
358 Magmatic Province: Palaeogeography, palaeoclimatology, palaeoecology, v. 441, p. 74–82,
359 <http://dx.doi.org/10.1016/j.palaeo.2015.04.011>.

360 Schlager, W., 2010, Ordered hierarchy versus scale invariance in sequence stratigraphy:
361 International Journal of Earth Sciences, v. 99, p. 139–151, doi:10.1007/s00531-009-0491-8.

362 Schlager, W., 1981, The paradox of drowned reefs and carbonate platforms: Geological Society
363 of America Bulletin, v. 92, p. 197–211.

364 Scotese, C.R., 2016, PALEOMAP PaleoAtlas for GPlates and the PaleoData Plotter Program
365 (PALEOMAP Project, 2016);, <https://www.earthbyte.org/paleomap-paleoatlas-for-gplates/>.

366 Scotese, C.R., and Wright, N., 2018, PALEOMAP Paleodigital Elevation Models (PaleoDEMS)

367 for the Phanerozoic (PALEOMAP Project, 2018):, <https://www.earthbyte.org/paleodem->
368 [resource-scotese-and-wright-2018/](https://www.earthbyte.org/paleodem-resource-scotese-and-wright-2018/).

369 Skelton, P.W., 2003, The operation of the major geological carbon sinks, *in* Skelton, P.W.,
370 Spicer, R.A., Kelley, S.P., and Gilmour, I. eds., *The Cretaceous World*, Cambridge
371 University Press, p. 259–266.

372 Skelton, P.W., and Gili, E., 2011, Rudists and carbonate platforms in the Aptian: a case study on
373 biotic interactions with ocean chemistry and climate: *Sedimentology*, v. 59, p. 81–117,
374 <http://doi.wiley.com/10.1111/j.1365-3091.2011.01292.x>.

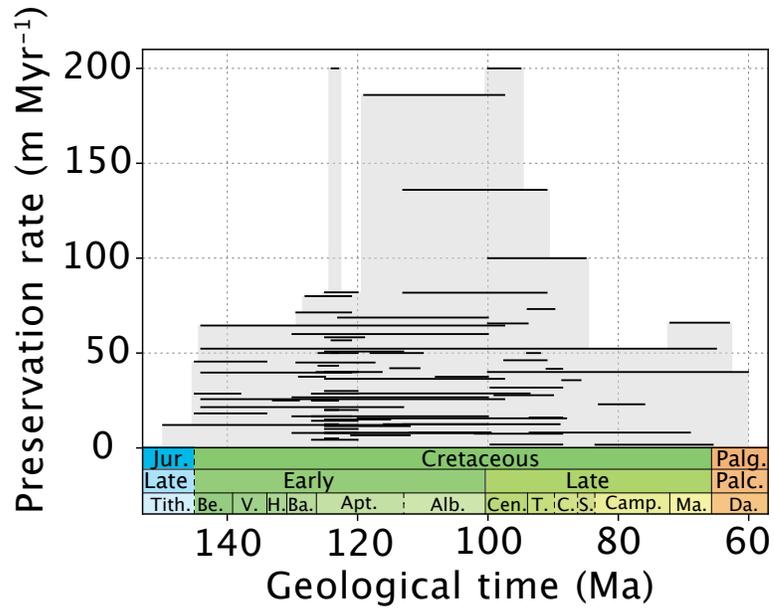
375 Strasser, A., and Samankassou, E., 2003, Carbonate sedimentation rates today and in the past:
376 Holocene of Florida Bay, Bahamas, and Bermuda vs. Upper Jurassic and lower cretaceous
377 of the Jura mountains (Switzerland and France): *Geologia Croatica*, v. 56, p. 1–18,
378 [doi:10.4154/GC.2003.01](https://doi.org/10.4154/GC.2003.01).

379 Veizer, J. et al., 1999, $^{87}\text{Sr}/^{86}\text{Sr}$, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ evolution of Phanerozoic seawater: *Chemical*
380 *Geology*, v. 161, p. 59–88, [doi:10.1016/S0009-2541\(99\)00081-9](https://doi.org/10.1016/S0009-2541(99)00081-9).

381 Whitten, E.H.T., 1976, Cretaceous phases of rapid sediment accumulation, continental shelf,
382 eastern USA: *Geology*, v. 4, p. 237–240, [doi:10.1130/0091-](https://doi.org/10.1130/0091-7613(1976)4<237:CPORSA>2.0.CO;2)
383 [7613\(1976\)4<237:CPORSA>2.0.CO;2](https://doi.org/10.1130/0091-7613(1976)4<237:CPORSA>2.0.CO;2).

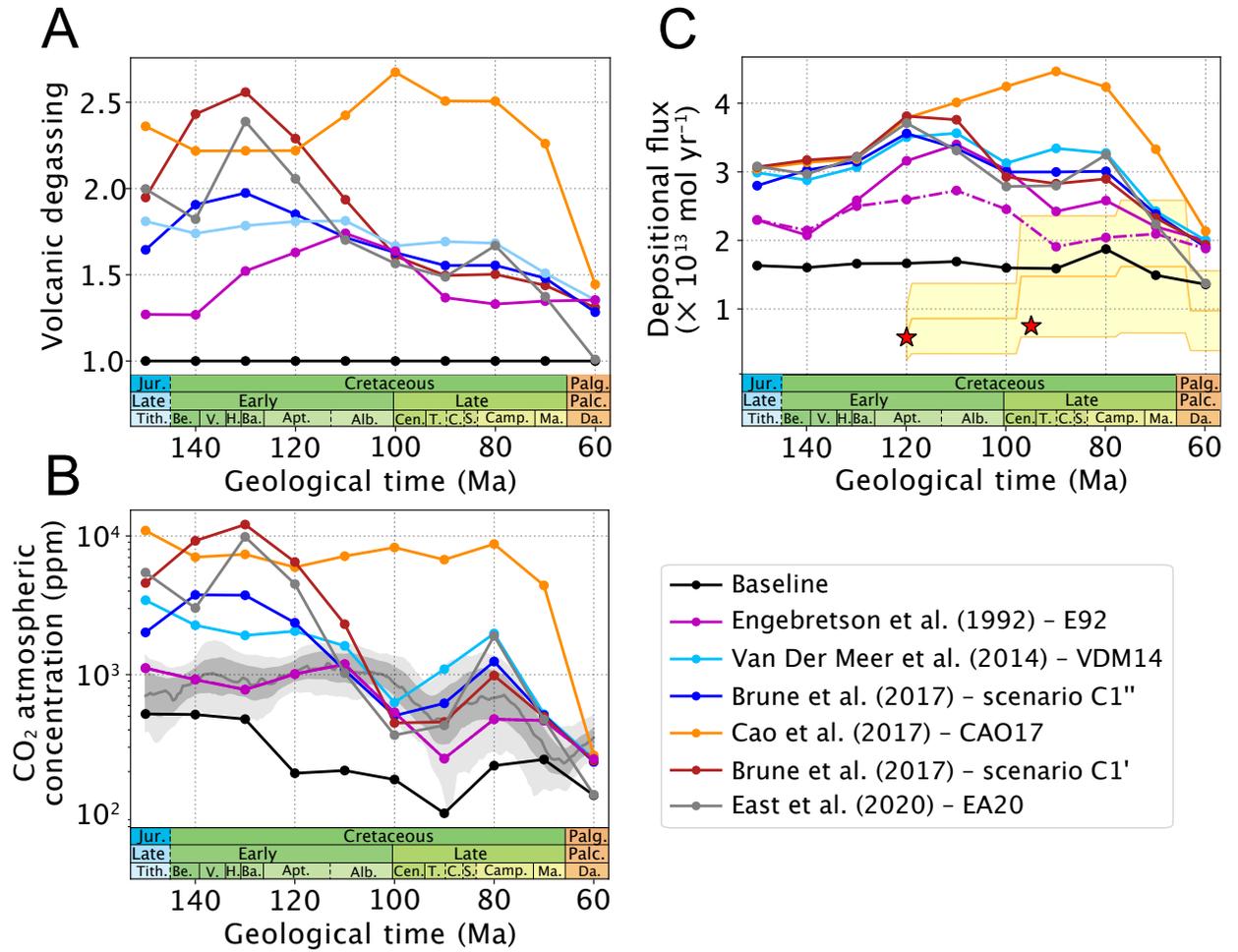
384

385 FIGURE CAPTIONS



386

387 Figure 1. Compilation of Cretaceous neritic carbonate preservation rates. The light grey
 388 envelope represents the maximum accumulation rates, which are likely to best represent neritic
 389 carbonate production in ancient times (see Methods). Stratigraphic boundaries after Gradstein et
 390 al. (2012).



391

392

393

394

395

396

397

398

399

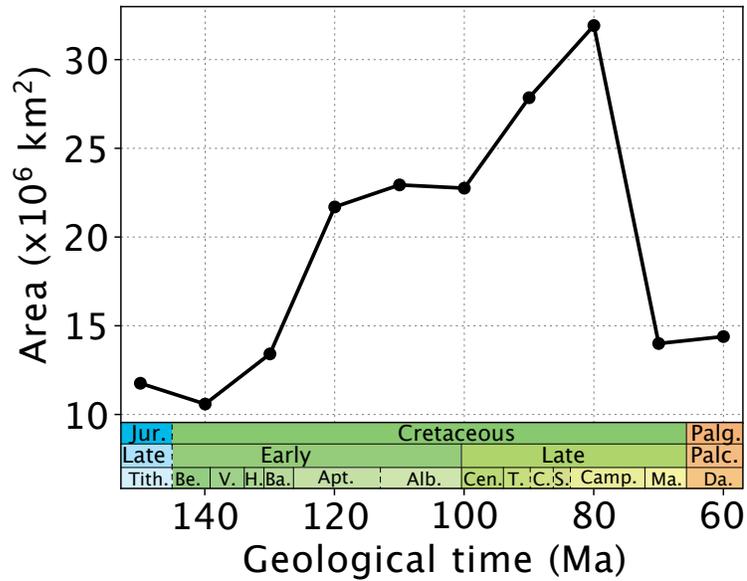
400

Figure 2. Model results. (A) Volcanic degassing scenarios used in the GEOCLIM model (ratio to present-day value), (B) predicted atmospheric CO₂ concentration and (C) predicted neritic carbonate total depositional flux. Black line represents the baseline runs with volcanic degassing set to the present-day level. Solid color lines represent volcanic degassing scenarios. For scenario E92, the simulation represented with the dashed line is identical to the one represented with the solid line except that it uses a constant area of the shallow-marine environments available to neritic carbonate deposition during the Cretaceous. In panel B, background grey shading represents the proxy-based pCO₂ reconstruction of Foster et al. (2017) with LOESS fit, 68 % and 95 % confidence intervals. In panel C, yellow shading and red stars

401 respectively represent the platform carbonate accumulation fluxes estimated by Dutkiewicz et al.
 402 (2019) and Skelton (2003) (see main text).

403

404



405

406 Figure 3. Area of the shallow-marine environments available to neritic carbonate
 407 deposition, derived from the paleogeographical reconstructions (see the Data Repository).

408

409 ¹GSA Data Repository item 201Xxxx, [Supplementary information, geological database,
 410 experimental setup and climate model runs (Sections DR1–DR2, Figures DR1–DR2, and Tables
 411 DR1–DR2)], is available online at www.geosociety.org/pubs/ft20XX.htm, or on request from
 412 editing@geosociety.org.