

1 **Title: Ocean Acidification and the Permo-Triassic Mass Extinction**

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18 **Abstract:** Ocean acidification triggered by Siberian Trap volcanism has been implicated as a kill
19 mechanism for the Permo-Triassic mass extinction, but evidence for an acidification event
20 remains inconclusive. To address this, we present a high resolution seawater pH record across
21 this interval, utilizing boron isotope data combined with a quantitative modeling approach. In the
22 latest Permian, the alkalinity of the ocean increased, priming the Earth system with a low level of
23 atmospheric CO₂ and a high ocean buffering capacity. The first phase of extinction was

24 coincident with a slow injection of isotopically light carbon into the atmosphere-ocean, but the
25 ocean was well-buffered such that ocean pH remained stable. During the second extinction pulse,
26 however, a rapid and large injection of carbon overwhelmed the buffering capacity of the ocean,
27 causing an abrupt and short-lived acidification event that drove the preferential loss of heavily
28 calcified marine biota.

29

30 **One Sentence Summary:** Ocean acidification caused the second phase of mass extinction in the
31 Permo-Triassic, due to a rapid and large injection of carbon which overwhelmed the buffering
32 capacity of the ocean.

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34 **Main Text:** The Permian Triassic Boundary (PTB) mass extinction, at ~ 252 Ma, represents the
35 most catastrophic loss of biodiversity in geological history, and played a major role in dictating
36 the subsequent evolution of modern ecosystems (1). The end-Permian extinction event spans ~60
37 kyrs (2) and can be resolved into two distinct marine extinction pulses, with the respective kill
38 mechanisms appearing to be ecologically selective (3). The first occurred in the latest Permian
39 (Extinction Pulse 1; EP1) and was followed by an interval of temporary recovery before the
40 second pulse (EP2) which occurred in the earliest Triassic. The direct cause of the mass
41 extinction is widely debated with a diverse range of overlapping mechanisms proposed,
42 including widespread water column anoxia (4), euxinia (5), global warming (6) and ocean
43 acidification (7).

44 Models of PTB ocean acidification suggest that a massive, and rapid, release of CO₂ from
45 Siberian Trap volcanism, acidified the ocean (7). Indirect evidence for acidification comes from
46 the interpretation of faunal turnover records (3, 8), potential dissolution surfaces (9) and Ca

47 isotope data (7). A rapid input of carbon is also potentially recorded in the negative carbon
48 isotope excursion (CIE) that characterizes the PTB (10, 11) . The interpretation of these records
49 is, however, debated (12), and of great importance to understanding the current threat of
50 anthropogenically-driven ocean acidification (11).

51 Here, we test the ocean acidification hypothesis by presenting a novel proxy record of
52 ocean pH across the PTB, using the boron isotope composition of marine carbonates ($\delta^{11}\text{B}_{\text{carb}}$)
53 (SM sections 2 to 4). We then employ a carbon cycle model (SM sections 3 to 5) to explore
54 ocean carbonate chemistry and pH scenarios that are consistent with our $\delta^{11}\text{B}$ data and published
55 records of carbon cycle disturbance and environmental conditions. The quantitative model uses
56 previous estimates for background Early Permian conditions (13) suggesting either high $p\text{CO}_2$
57 (CO_2Hi : ~10 PAL, pH ~7.5 $\delta^{11}\text{B}_{\text{SW}}$ ~36.8‰), or low $p\text{CO}_2$ (CO_2Lo : ~3 PAL, pH ~8, $\delta^{11}\text{B}_{\text{SW}}$
58 ~34‰) (see SM sections 5 and 6 for further details). Through this combined geochemical,
59 geological and modelling approach we are able to produce an envelope that encompasses the
60 most realistic range in pH, which then allows us to resolve three distinct chronological phases of
61 carbon cycle perturbation, each with very different environmental consequences for the Late
62 Permian-Early Triassic Earth system.

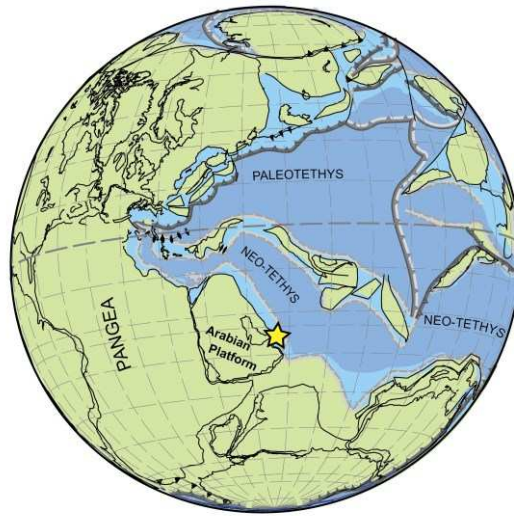
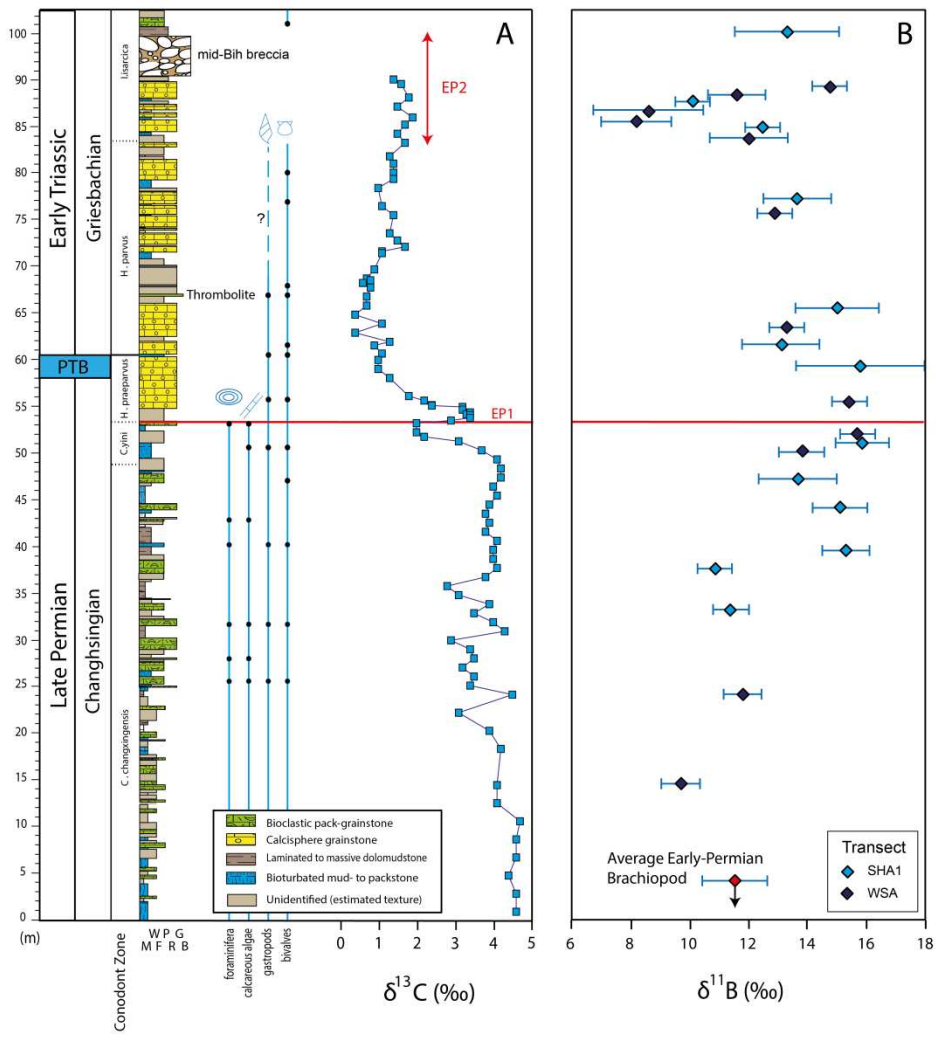


Figure 1

63

Figure 2



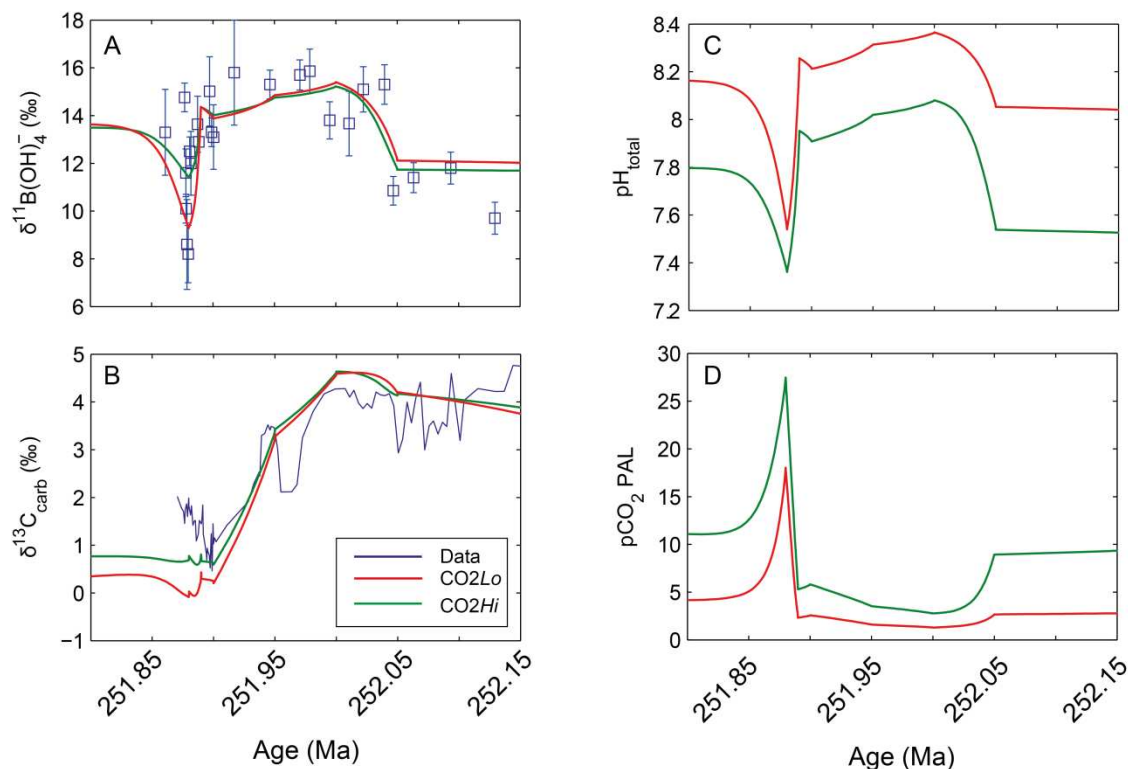
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65 We analyzed boron isotope data from two complementary transects in a shallow marine,
66 open water carbonate succession from the United Arab Emirates (U.A.E.), where depositional
67 facies and $\delta^{13}\text{C}_{\text{carb}}$ are well constrained (14) (see SM sections 1 and 2). During the Permian-
68 Triassic the U.A.E formed an expansive carbonate platform that remained connected to the
69 central Neo-Tethyan Ocean (15) (Fig 1). The run-up to PTB in the Tethys is characterized by
70 two negative $\delta^{13}\text{C}$ excursions interrupted by a short-term positive event (10). There is no
71 consensus as to the cause of this rebound event and so we focus on the broader $\delta^{13}\text{C}$ trend. Our
72 $\delta^{13}\text{C}$ transect (Fig. 2) starts in the Changhsingian with a gradual decreasing trend, interrupted by
73 the first negative shift in $\delta^{13}\text{C}$ at EP1 (at 53 m or ~ 251.96 Ma, Fig. 2). This is followed by the
74 minor positive ‘rebound’ event (at 54 m or ~ 251.95 Ma, Fig. 2) prior to the minima of the second
75 phase of the negative CIE (58-60 m or ~ 251.92 Ma, Figs. 2 and 3) that marks the PTB itself.
76 After the CIE minimum, $\delta^{13}\text{C}$ gradually increases to $\sim 1.8\text{‰}$ and remains relatively stable during
77 earliest Triassic and across EP2.

78 Our boron isotope transect shows a quite different pattern. $\delta^{11}\text{B}$ is persistently low (Fig. 2),
79 at the start of our record during the late-Changhsingian (Late Permian), with an average of $10.9 \pm$
80 0.9‰ (1σ). This is in agreement with $\delta^{11}\text{B}$ values (average of $10.6 \pm 0.6\text{‰}$, 1σ) reported for
81 early-Permian brachiopods (16). Further up section (at ~ 40 m or ~ 252.04 Ma, Fig. 2), there is a
82 stepped increase in $\delta^{11}\text{B}$ to 15.3‰ , and by implication an increase in ocean pH of ~ 0.4 - 0.5 pH
83 units (Fig. 3). Then $\delta^{11}\text{B}$ values remain relatively stable, scattering around 14.8‰ (± 1.0 , 1σ) and
84 implying variations within 0.1 - 0.2 pH units, into the Early Griesbachian (Early Triassic) and
85 hence across EP1 and the period of carbon cycle disturbance (Figs. 2 and 3).

86 After $\delta^{13}\text{C}$ had increased and stabilized (at ~ 85 m or ~ 251.88 Ma, Fig. 2), $\delta^{11}\text{B}$ begins to
87 decrease rapidly to 8.2‰ , implying a sharp drop in pH of ~ 0.6 - 0.7 pH units. The $\delta^{11}\text{B}$ minimum

Figure 3



88 is coincident with the interval identified as EP2. This ocean acidification event is short-lived
 89 (~10 kyrs) and $\delta^{11}\text{B}$ values quickly recover toward the more alkaline values evident during EP1
 90 (average of ~14‰).

91 The initial rise in ocean pH of ~0.4-0.5 units during the Late Permian (Fig. 3) suggests
 92 a large increase in carbonate alkalinity (17). We are able to simulate the observed rise in $\delta^{11}\text{B}$
 93 and pH through different model combinations of increasing silicate weathering, increased pyrite
 94 deposition (18), an increase in carbonate weathering and a decrease in shallow marine carbonate
 95 depositional area (see SM section 6.1). Both silicate weathering and pyrite deposition result in a
 96 large drop in pCO_2 (and temperature) for a given increase in pH and saturation state (Ω). There is
 97 no evidence for a large drop in pCO_2 , and independent proxy data indicate only a minor
 98 temperature decrease of a few degrees C during the Changhsingian (19), suggesting that these
 99 mechanisms alone cannot explain the pH increase (see sensitivity tests in SM section 6.1).

100 Conversely, an increase in carbonate input or a reduction in rates of carbonate deposition both
101 result in increases in Ω , with a greater impact on pH per unit decrease in $p\text{CO}_2$ and temperature
102 (see Fig. S6).

103 This suggested decrease in carbonate sedimentation is consistent with the decrease in
104 depositional shelf area that occurred due to the 2nd order regression of the Late Permian (20).
105 With the added expansion of anoxia into shelf environments (21) this would effectively create
106 both bottom-up and top-down pressures to reduce the area of potential carbonate sedimentation.
107 Sea level fall also exposed carbonates to weathering (20), which would have further augmented
108 the alkalinity influx. The pH increase event supports the CO_2Lo initialisation scenario ($\text{CO}_2 \sim 3$
109 PAL, pH ~ 8 , $\delta^{11}\text{B}_{\text{SW}} \sim 34\%$) as the simulated CO_2 and temperature decrease is much reduced,
110 and therefore more consistent with independent proxy data (22), compared to CO_2Hi ($\text{CO}_2 \sim 10$
111 PAL, pH ~ 7.5 , $\delta^{11}\text{B}_{\text{SW}} \sim 36.8\%$) (Fig. 3D).

112 Prior to EP1, $\delta^{13}\text{C}_{\text{carb}}$ values begin to decrease before reaching the minimum of the
113 globally recognized negative CIE at the PTB (Fig. 2). At this time both $\delta^{11}\text{B}$ and ocean pH
114 remained stable. Hypotheses to explain the negative CIE require the input of isotopically light
115 carbon, such as from volcanism (14, 23) with the assimilation of very light organic carbon from
116 the surrounding host rock (24), methane destabilization (22), collapse of the biological pump
117 (15), and/or a decrease in the burial of terrestrial carbon (16). We can simulate the observed drop
118 in $\delta^{13}\text{C}_{\text{carb}}$, whilst remaining within the uncertainty of the $\delta^{11}\text{B}$ data (Fig. 3), by combining a
119 cessation of terrestrial carbon burial with a relatively slow (50 kyr) carbon injection from any of
120 the above sources (see section S6.2.4, Fig S8). A small source of methane (3.2×10^{17} mol C with
121 $\delta^{13}\text{C} = -50\%$) gives the least change in $\delta^{11}\text{B}$ and pH, whilst either a larger source of organic

122 carbon ($\sim 6.5 \times 10^{17}$ mol C with $\delta^{13}\text{C} = -25\%$) or a mixture of mantle and lighter carbon sources
123 ($\sim 1.3 \times 10^{18}$ mol C with $\delta^{13}\text{C} = -12.5\%$) are still within the measured uncertainty in $\delta^{11}\text{B}$.

124 This relatively slow carbon addition minimises the tendency for a transient decline in
125 surface ocean pH in an ocean that was already primed with a high Ω and hence high buffering
126 capacity from the Late Permian. The global presence of microbial and abiotic carbonate fabrics
127 after EP1 (Fig. 2) (25) are indicative that this high Ω was maintained across the CIE. The carbon
128 injection triggers an increase in $p\text{CO}_2$, temperature and silicate weathering, thereby creating an
129 additional counterbalancing alkalinity flux. This is consistent with independent proxy data (6).
130 The alkalinity source may have been further increased through soil loss (26), the emplacement of
131 easily-weathered Siberian Trap basalt, or the impact of acid rain (27) that would have increased
132 weathering efficiency.

133 The negative $\delta^{11}\text{B}_{\text{carb}}$ excursion at 251.88 Ma represents a calculated pH decrease of up to
134 0.7 pH units. It coincides with the second pulse of the extinction (Fig. 2), which preferentially
135 affected the heavily calcifying, physiologically un-buffered and sessile organisms (3). This was
136 also accompanied by the temporary loss of abiotic and microbial carbonates throughout the
137 Tethys (28, 29) thereby suggesting a coeval decrease in Ω (30). To overwhelm the buffering
138 capacity of the ocean and decrease pH in this way requires a second, more abrupt injection of
139 carbon to the atmosphere, yet remarkably, the acidification event occurs after the decline in $\delta^{13}\text{C}$,
140 when $\delta^{13}\text{C}$ has rebounded somewhat and is essentially stable (Fig. 2).

141 Unlike the first carbon injection, the lack of change in $\delta^{13}\text{C}$ at this time rules out very
142 ^{13}C -depleted carbon sources, because no counterbalancing strongly ^{13}C -enriched source exists.
143 Instead, it requires a carbon source near $\sim 0\%$. A plausible scenario for this is the decarbonation
144 of overlying carbonate host rock, into which the Siberian Traps intruded (24) or the direct

145 assimilation of carbonates and evaporites into the melt (31). Host carbonates would have had
146 $\delta^{13}\text{C} \sim +2\text{-}4\text{‰}$, which when mixed with mantle carbon ($\sim -5\text{‰}$), potentially produces a source
147 near 0‰. We can simulate the sharp drop in pH and stable $\delta^{13}\text{C}$ values (Fig. 3) through a large
148 and rapid carbon release of 2×10^{18} mol C over 10 kyr (Fig S8). This second rapid carbon release
149 produces a sharp rise in $p\text{CO}_2$ to ~ 20 PAL and warming of $\sim 15^\circ\text{C}$, consistent with the
150 observation of peak temperatures after EP1 (26). Initialization of the carbon cycle model under
151 CO_2Hi cannot generate the magnitude of $\delta^{11}\text{B}$ drop (Fig. 3A) because the non-linear relation
152 between pH and $\delta^{11}\text{B}$ fractionation sets a lower limit of $\delta^{11}\text{B}$ at $\sim 10\text{‰}$ in this case (Fig. S3). Thus
153 low initial CO_2 of ~ 3 PAL in the late Permian (CO_2Lo) is more consistent with the data
154 presented here.

155 The documented acidification event lasted for only ~ 10 kyrs. This time span is consistent
156 with the modelled timescale required to replenish the ocean with alkalinity, as carbonate
157 deposition is reduced and weathering is increased under higher $p\text{CO}_2$ and global temperatures.
158 Increased silicate weathering rates drive further CO_2 drawdown resulting in stabilization (Fig.
159 S7). High global temperature (6) and increased silicate weathering are consistent with a sudden
160 increase in both $^{87}\text{Sr}/^{86}\text{Sr}$ (32) and sedimentation rates (26) in the Griesbachian.

161 The Permo-Triassic transition was a time of extreme environmental change, and our
162 combined data and modeling approach falsifies several mechanisms for the changes observed.
163 Whilst the coincident stresses of anoxia, increasing temperatures, and ecosystem restructuring
164 were important during this interval, the $\delta^{11}\text{B}$ excursion strongly suggests that widespread ocean
165 acidification was the driver of the second pulse of the mass extinction, and thus ultimately ended
166 the temporary recovery (3) from the first extinction pulse. The carbon release required to drive
167 the observed acidification event occurred at a rate comparable to the current anthropogenic

168 perturbation, but far exceeded it in expected magnitude (33). We show that such a rapid rate is
169 critical to causing the combined synchronous decrease in both pH and saturation state that
170 defines an ocean acidification event (11).

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173 **References and Notes**

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465
466 **Fig. 1:** Paleogeographic reconstruction for the Late Permian denoting the studied section Wadi
467 Bih, in the Musandam Mountains of U.A.E that formed an extensive carbonate platform in the
468 Neo-Tethyan Ocean. Modified from (14).

469
470 **Fig. 2:** Carbon cycle dynamics, macrofauna distribution (34), and Extinction Phase 1 (EP1) and
471 Extinction Phase 2 interval (EP2) across the Permian Triassic Boundary for Wadi Bih, A:
472 shallow water $\delta^{13}\text{C}$ record (14); B: Boron isotope ($\delta^{11}\text{B}$) record (propagated uncertainty given as
473 $2\sigma_f$) and average Early Permian brachiopod value ($n=5$) (16). Blue symbols are for SHA1
474 transect, black symbols are for WSA transect. The two transects are separated by $\sim 1\text{km}$ and
475 facies are laterally continuous. Only *H. parvus* has been found so far in this section and the
476 conodont zones with dashed line are identified from the $\delta^{13}\text{C}$ record (34-36).

477
478 **Fig. 3:** Model results of carbon cycle parameters for the two end-member CO_2 scenarios (CO_2Hi
479 and CO_2Lo). A - Model reproduced $\delta^{11}\text{B}$ vs data. B - Modelled $\delta^{13}\text{C}$ vs data. C - Modelled pH

480 envelope incorporating uncertainty of $\delta^{11}\text{B}_{\text{SW}}$ and dynamic temperatures. D – calculated

481 atmospheric CO_2 . See SM section 1.2 for details on age model.

482

483

484 **Supplementary Materials:**

485 Materials and Methods

486 Figures S1-S9

487 Tables S1-S10

488 References (37-96)

489

490 **Supplementary Materials**

491 **1. Materials and Methods**

492 **1.1 Sample Section**

493 Carbonate samples were obtained from a shallow-marine platform section at Wadi Bih on the
494 Musandam Peninsula, United Arab Emirates (UAE). GPS co-ordinate for main transect in Wadi
495 Shahha are N 025° 50' 31.7", E 056° 06' 41.7". This is a near-continuous and exclusively
496 shallow marine carbonate succession from the late Permian and entire Early Triassic that allows
497 analysis of a high-resolution signature in the absence of significant depth-dependent or
498 lithological controls (14). Late Permian facies are predominantly intertidal mud-and-wackestones
499 and open lagoon wackestones, occasional windward and leeward shoals of bioclastic pack-and-
500 grainstones (34). The PTB and Early Triassic are regionally represented by ooidal grainstones
501 and a thrombolite horizon (14, 34). Samples were taken from two transects Wadi Shahha (WSA)
502 and Sha (SHA1), ~1000 m apart.. Facies are laterally continuous and carbon isotopes were used
503 to ensure careful integration of the two datasets. For this study, we have analyzed 25 carbonate
504 samples in total, taken in approximately 1 to 10 m scale intervals and covering a total thickness
505 of 98 m, starting in the mid-Changhsingian (Permian), going up into the mid-Griesbachian
506 (Triassic) and bracketing the complete Permian Triassic Boundary (PTB) extinction event (Table
507 S1).

Table S1: Boron, carbon and oxygen isotope data from Wadi Bih; U.A.E.; pH values are derived from a simplified and non-dynamic calculation using the $\delta^{11}\text{B}_{\text{SW}}$ range given by the model calculation and a constant pK_B for 25°C, 35 psu, 0 dbar.

Sample	Height (m)	Lithology	Age (Ma)*	$\delta^{13}\text{C}$ (‰)	$\delta^{18}\text{O}$ (‰)	$\delta^{11}\text{B}$ (‰) ⁺	2 σ (‰) [#]	pH	
								$\delta^{11}\text{B}_{\text{SW}}$ 34‰	$\delta^{11}\text{B}_{\text{SW}}$ 36.8‰
Wadi Bih (25° 50' 31.2"N 56° 06' 41.7"E)									
WSA54	14.0	micritic mudstone	252.13	4.20	-2.40	9.7(3)	0.7	7.7	6.7
WSA49	24.0	micritic mudstone	252.09	4.60	-0.50	11.8(3)	0.7	8.0	7.6
SHA1 1	32.4	dolograinstone	252.06	4.23	-3.19	11.4(2)	0.6	7.9	7.5
SHA1 3	37.0	grainstone	252.05	4.22	-2.62	10.8(0)	0.6	7.9	7.4
SHA1 4	39.0	micritic mudstone	252.04	4.26	-2.51	15.3(6)	0.8	8.3	8.1
SHA1 7	43.8	bioclastic packstone	252.02	3.63	-2.65	15.1(7)	1.0	8.3	8.0
SHA1 8	47.0	bioclastic packstone	252.01			13.7(12)	1.4	8.2	7.9
WSA22a	50.0	micritic mudstone	252.00	3.80	-3.50	13.8(5)	0.8	8.2	7.9
SHA1 11	51.0	bioturbated mudston	251.98			15.9(7)	0.9	8.3	8.1
WSA21	52.0	micritic mudstone	251.97	2.10	-4.00	15.7(2)	0.6	8.3	8.1
WSA 17	55.0	micritic mudstone	251.95	2.54	-3.65	15.3(2)	0.6	8.3	8.1
SHA1 15	59.6	calcisphere grainston	251.92	0.52	-3.15	15.8(21)	2.2	8.3	8.1
SHA1 17	62.0	calcisphere grainston	251.90			13.1(12)	1.4	8.1	7.8
WSA12A	64.0	micritic mudstone	251.90	0.50	-3.40	13.3(0)	0.6	8.1	7.8
SHA1 20	66.0	calcisphere grainston	251.90			15.0(13)	1.5	8.3	8.0
WSA5A	77.0	micritic mudstone	251.89	1.50	-3.20	12.9(0)	0.6	8.1	7.8
SHA1 26	78.2	calcisphere grainston	251.89	1.21	-2.55	13.6(10)	1.2	8.2	7.9
WSA3	85.0	micritic mudstone	251.88	1.60	-2.60	12.0(12)	1.3	8.0	7.6
SHA1 30	86.2	calcisphere grainston	251.88			12.5(6)	0.6	8.1	7.7
WSA 2A	86.8	micritic mudstone	251.88	1.99	-2.96	8.2(10)	1.2	7.4	-
WSA2	88.0	micritic mudstone	251.88	1.60	-2.94	8.6(18)	1.9	7.5	<7,0
WSA 1B	89.0	micritic mudstone	251.88	1.87	-3.26	10.1(2)	0.6	7.8	7.0
SHA1 32	90.2	calcisphere grainston	251.88	1.39	-3.40	11.6(8)	1.0	8.0	7.6
WSA1	91.0	micritic mudstone	251.88	1.46	-3.74	14.8(0)	0.6	8.2	8.0
SHA1 36	102.0	bioclastic packstone	251.86	1.79	-2.00	13.3(17)	1.8	8.1	7.8

* Ages are calculated based on tie points from Burgess et al.(2), see section S1.2

⁺ external uncertainty for B isotope data are valid for the last digits (2 σ)

[#] propagated uncertainties for B isotope data

508

509

510 1.2 Age Model

511 The age model for this study was based on the most recent findings of Burgess et al., (2) (see
512 Table S1). The age tie points are 252.10 Ma for 20 m (Fig. 2) thought to be equivalent to Bed 22
513 in Meishan from the $\delta^{13}\text{C}$ record; 252.00 Ma for the decline in $\delta^{13}\text{C}$ at 49 m and 251.90 Ma for
514 the first occurrence of *H. parvus* at 61m. EP2 is known to occur at the end of the *I. staeschi* zone,
515 Bed 28 Meishan (3), which is not recorded in the Neo-Tethys but is equivalent to the carbon

516 isotope rise between the *H. Parvus* and *I. isarcica* zones. In Wadi Bih we see the loss of
517 gastropods and bivalves at this point (34), however the full resolution of the second extinction
518 phase has not been confirmed at this locality. Therefore we identify a 10 m interval for EP2. In
519 order to complete the age model we set the mid-point of EP2 as 251.88 Ma based on the absolute
520 age for EP2 in Meishan (2).

521

522 **1.3 Analytical methods**

523 Carbon and oxygen isotope ratios were determined at Edinburgh University and University of
524 Graz using the preparation technique and measurement routine as detailed in (19). All carbonate
525 isotopic values are quoted in the conventional δ per mil (‰) notation relative to VPDB.
526 Calibration to international reference material was through NBS 19 and the reproducibility of
527 replicate analyses for reference material, standards (in-house) and carbonate samples was better
528 than ± 0.05 ‰ for $\delta^{13}\text{C}$ and ± 0.1 ‰ for $\delta^{18}\text{O}$ at one standard deviation. For a detailed discussion
529 of the C and O isotopic data see (19).

530 Boron isotope ratios were determined in the isotope geochemistry laboratory at the
531 University of Bremen (Department of Geosciences and MARUM-Center for Marine
532 Environmental Sciences) using a ThermoFisher Scientific TRITON Plus Thermal Ionization
533 Mass Spectrometer. The $n(^{11}\text{B})/n(^{10}\text{B})$ measurements were performed using negative thermal
534 ionization mass spectrometry (N-TIMS) following the method detailed in (37). For analyses, 10
535 mg of the sample powder was dissolved in 100 μl 1 N HCl for 24 h at 20°C and subsequently
536 centrifuged. 1 μl boron-free seawater emitter (38) together with 1 μl of the sample solution
537 containing ~ 1 ng B were placed and dried on a degassed Re single filament. Analyses were
538 carried out at filament temperatures of 970°-1050°C. Boron isotopes were registered as BO_2^-

539 complexes on masses 42 and 43, and measurements were carried out at ion beam intensity of ~
540 10 pA on mass 43. Each sample measurement involved up to 200 blocks with 10 cycles each,
541 taking about 2 hours of data acquisition. B isotope ratios are given relative to NIST SRM 951 in
542 the conventional $\delta^{11}\text{B}$ (‰) notation. The external reproducibility of the certified reference
543 material NIST SRM 951 showed an $n(^{11}\text{B})/n(^{10}\text{B})$ ratio of 4.0065 ± 0.0017 ($2\sigma = 0.42\%$, $n=23$)
544 over a period of 6 month of sample analyses. The long term (18 month) reproducibility is
545 4.0066 ± 0.0018 ($2\sigma = 0.44\%$, $n=58$). The $n(^{11}\text{B})/n(^{10}\text{B})$ ratio of the reference material for each
546 analytical session was reproduced better than 0.6% (2σ). In addition to the NIST material, the
547 standard material M93-TB-FC-1, a *Porites* coral with a published value of $24.8\pm 0.4\%$ ($2\sigma_{\text{mean}}$),
548 as determined by different multicollector techniques (39) was also regularly analysed. The coral
549 replicates gave $\delta^{11}\text{B}$ values of $24.1\pm 0.7\%$ (2σ , $n=16$). Each sample solution was as a minimum
550 run in full duplicate and the uncertainty of the reference material was propagated into the
551 external uncertainty (2σ) on the isotope ratio of the sample ($2\sigma_f$; Table S1). For further
552 information on the preparation and analytical methods see (40).

553

554 **2. Sample Selection**

555 A critical issue in using the B isotope composition of Permo-Triassic carbonate rocks for
556 reconstruction of ocean pH is the potential overprinting of the original isotope signal via
557 alteration. To obtain high-quality samples in which primary B isotope signals are still preserved,
558 we applied the sample selection procedures successfully used on Neoproterozoic carbonate rocks
559 to reconstruct Cryogenian and Ediacaran ocean pH variation (37, 40, 41). In brief, samples
560 selected for isotope analyses were screened macroscopic (in the field), microscopic (scanning
561 electron microscope) and geochemically (carbon and oxygen isotopes), and selected because of

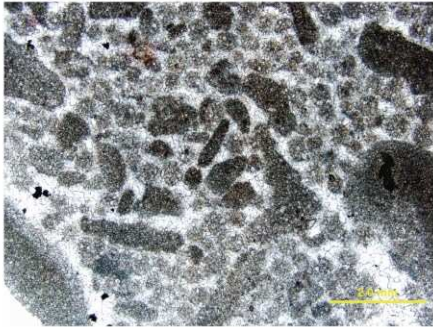
562 their uniformity in texture, absence of late stage secondary alteration and no correlation between
563 carbon and boron with oxygen isotopic signatures (Figs. S1 and S2).

564 The PTB interval itself was thought to occur in a regional cross-bedded ooid grainstone,
565 however, reconsideration of published thin-section images (34) (Fig S1 samples from WSA
566 transect, 0611300847 and 0611181243) demonstrate these are in fact microspheres (a.k.a
567 calcispheres). This fabric is unusual for the Phanerozoic but appears to represent supersaturated
568 conditions and these are thought to be primary spar precipitates (42).

569 To gain the highest potential for recording a primary boron signature we microdrilled
570 areas of pure micrite from micritic mudstones and micritised calcisphere grainstones where there
571 was no micro-optical evidence for veining, fracturing, clay minerals or bioclastic material.
572 Samples containing evidence of late stage secondary alteration or recrystallization, i.e. any
573 spar/microspar, were discarded. Potential analyses of disseminated detrital material in the
574 carbonates was also checked through elevated Al, Si and Ba concentrations and showed either no
575 or only minor dissemination of clay. The low availability of bioclastic material in this interval,
576 due to the extinction, argues against the potential of trends being caused by vital effects. In
577 particular the calcisphere grainstones of the PTB interval are classed as an abiotic carbonate that
578 provides an excellent opportunity to record seawater $\delta^{11}\text{B}$ in the absence of vital effects.

579

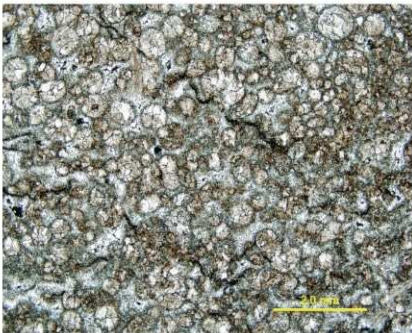
0611301104 (87m)



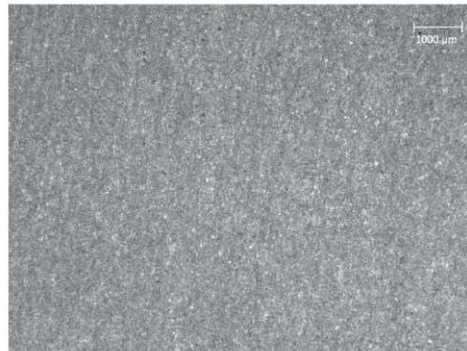
WSA1 (91m)



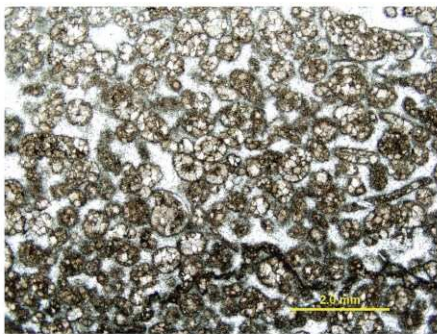
0611300847 (77m)



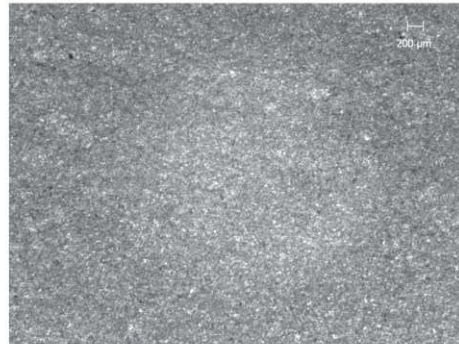
WSA2 (86m)



0611181243 (56m)

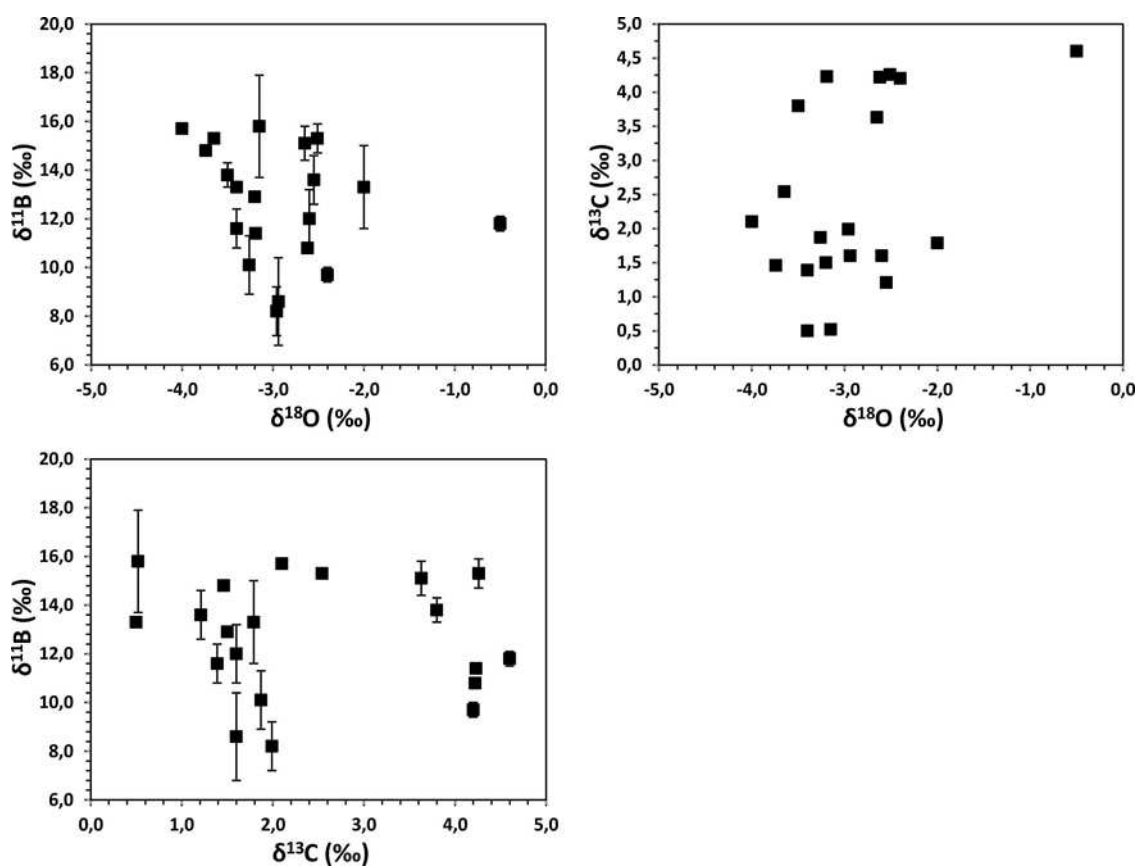


WSA3 (85m)



580

581 Fig S1: Examples of thin section images for the Wadi Bih section, WSA transect. WSA samples
582 demonstrate homogeneous micritic texture of samples used for $\delta^{11}\text{B}$ analysis. 06- samples are from
583 Maurer et al., (34) illustrating grainstones. The lack of compaction of grains and pore filling equant
584 spar cements indicate very early/syn-sedimentary cementation. This fabric is consistent with
585 indicators for supersaturated conditions in this interval, including thrombolites and flat pebble
586 conglomerates, and suggests preservation of the original depositional fabric. Sample heights (m) as
587 in Fig. 2 and Table S1.



588

589 Fig S2: Cross-plots of boron, carbon, and oxygen isotope data for the marine carbonate samples
 590 from the Wadi Bih section. No significant correlation can be observed.

591

592 **2.1 Diagenetic Influences**

593 The PTB interval grainstone is composed of microspheres (a.k.a calcispheres). Microspheres are
 594 micritized and cemented with coarse equant spar cement (Fig S1). The lack of compaction of the
 595 microspheres suggests that the spar cement was extremely early and probably syn-sedimentary.

596 This is a feature of other grainstones throughout this interval (WSA transect, sample
 597 0611301104). Critically the presence of such well cemented fabrics in the grainstones of this
 598 interval would have prevented later stage fluid migration, and hence any late stage
 599 dolomitization. The total negative boron isotope excursion at EP2 is recorded across a number of
 600 fabrics and facies, including micrite and the early cemented grainstones. Petrographic analysis

601 shows the extreme of the acidification event is recorded in fine-grained micrite where textures
602 are very homogenous samples WSA 1, 1B, 2, and 2A (Fig. S1).

603 Dolomitization in the Wadi Bih section is observed (14), and occurred in two phases
604 where the first phase was fabric retentive syn-sedimentary dolomitization. The second phase of
605 dolomitization leads to the recrystallization of facies where near depositional porosity and
606 permeability had been maintained, i.e. preferentially in non-cemented grainstones. Neither of
607 these phases is classed as deep burial dolomitization (as confirmed by petrography in Figure S1
608 and the $\delta^{18}\text{O}$ data, Table S1). In the Musandam Peninsula the impact of dolomitization on carbon
609 and oxygen isotopes can be seen only in the close proximity to regional faults that provided high
610 permeability pathways for late burial dolomitizing fluids (43), which were avoided for this study.

611 Diagenetic effects were also checked with geochemical tests for selected trace element
612 analyses, using secondary ionisation mass spectrometer Cameca ims 4f at the University of
613 Edinburgh. Diagenetic alteration was explored via Mn/Sr, which is often used as a geochemical
614 indicator of alteration and meteoric diagenesis in ancient carbonates (44). While low Mn/Sr has
615 been found in ancient carbonates where there is clear optical evidence for recrystallization (45),
616 high Mn/Sr may indicate carbonate precipitation from anoxic waters. In the selected samples,
617 Mn concentration is $< 300 \mu\text{g g}^{-1}$, Sr concentration ranges from 140 to 1400 $\mu\text{g g}^{-1}$, and the
618 Mn/Sr ratio is low (< 2), so suggesting no influence of meteoric fluids. The boron concentration
619 for the carbonate rocks is, on average, $1 \mu\text{g g}^{-1}$ and ranges between 0.2 and $1.7 \mu\text{g g}^{-1}$.

620 Post-depositional alteration, especially meteoric diagenesis and recrystallization, is
621 assumed to decrease the isotopic composition of oxygen, boron and carbon isotopes (e.g. 46-48).
622 Our selected carbonate samples show the distinct carbon isotope trend found at every PTB
623 section globally and $\delta^{18}\text{O}$ values ranging between -0.5 and -4.0‰, indicating no significant deep

624 burial alteration. All carbon and oxygen isotope data are given in Table S1. No statistically
625 significant correlation between carbon, oxygen and boron isotope data can be observed in our
626 data (Fig. S2).

627 The replication of the $\delta^{11}\text{B}$ across two complementary transects further support the
628 preservation of a primary seawater isotope signature as late stage dolomitization is expected to
629 be highly laterally variable and disrupt the smooth trends seen in the data.

630

631 **3. Support for a primary Boron isotope signature**

632 ***3.1 Published $\delta^{11}\text{B}$ data***

633 So far, $\delta^{11}\text{B}$ data for Triassic carbonates have not been published and only limited $\delta^{11}\text{B}$ data,
634 based on low-Mg calcite brachiopod calcite from Oman (Saiwan Fm.), are available for the
635 Permian (Late Sakmarian). Keeping in mind the modern residence time of boron ($T \sim 14$ to 20 Ma
636 (47, 49), it is important to note that the range in $\delta^{11}\text{B}$ values ($10.9 \pm 0.9\%$ (1σ , $n=4$), 252.05-
637 252.15 Ma, mid-Changhsingian) overlap with the range in $\delta^{11}\text{B}$ values between 10.1 and 11.7‰
638 of the 285 Ma old brachiopod carbonates (average $10.6 \pm 0.0\%$, 1σ , $n=5$, (16)), which support the
639 primary nature of our boron isotope data.

640

641 ***3.2 Environmental controls on the boron isotope composition***

642 The speciation of boron in seawater is predominantly pH dependent with minor controls from
643 temperature, salinity and pressure (50). In our model we consider a dynamic temperature effect
644 on pK_{B} to calculate the pH of the seawater from the boron isotope composition preserved in
645 Permo-Triassic carbonate (see SM 4 and 5) however below we evaluate geological evidence to
646 consider whether these variables could drive the trends seen in our data.

647 3.2.1 Temperature

648 Oxygen isotope data preserved in conodont apatite and low-Mg calcite brachiopods from
649 Permian-Triassic sections in Iran document tropical sea surface temperatures (SST) of 27–33°C
650 during the Changhsingian with a negative shift in $\delta^{18}\text{O}$ starting at the extinction horizon,
651 translating into a warming of SSTs to well over 35°C in the mid-Griesbachian (51). The results
652 are consistent with SSTs of the South Chinese sections, where a rapid increase in seawater
653 temperature from about 25° to 36°C across the Permian-Triassic section was proposed based on
654 conodont apatite (6). Since the dissociation constant of boric acid (pK_B) decreases with
655 increasing temperature (50), the proposed increase in SST could drive an increase in the B
656 isotope composition of the carbonates precipitated at the time of global warming and imply an
657 increase in ocean pH, even if ocean pH stayed stable. Using the pK_B data from (50), the potential
658 increase in SST at a given pH of 8 would drive an increase in $\delta^{11}\text{B}$ of the carbonates by 2‰ and
659 generate an artificial increase of < 0.2 pH units. Consequently, it could be argued that our
660 calculated pH variations are in essence the result of an increase in sea surface temperatures.
661 However, the oxygen isotope pattern (51) shows a continuous increase in temperatures across the
662 Permian-Triassic boundary and a thermal maximum in the mid-Griesbachian (~251.85 Ma).
663 Instead our data show a sharp increase in $\delta^{11}\text{B}$ and ocean pH in the Late Permian, well before the
664 increase in temperatures at EP1. The slight variability in $\delta^{11}\text{B}$ above the PTB, however, could be
665 driven by temperature variations.

666

667

668

669

670 *3.2.2 Salinity*

671 Similar to temperature, the pK_B is sensitive to salinity and increases with decreasing salinity (52).
672 To drive significant salinity-associated isotope variations across the Permo-Triassic boundary,
673 drastic changes in salinity (e.g. down to 25 psu compared to 35 psu) would be necessary.
674 However, even if modelled sea surface salinity values for the late Permian indicate generally
675 more saline conditions than present day (52), sea surface salinity values for e.g. Iran and South
676 China still range between 34 and 35 psu (51) and will not have a significant effect on our pH
677 calculation. These samples come from an exclusively shallow water environment far from any
678 freshwater input, and so decreases in salinity would not affect our pH interpretations. The
679 depositional environment here ranges between open water ooid shoals to open and restricted
680 lagoon settings, which results in cyclic deposition on the scale of 5m. A shift toward more
681 restricted settings could drive increases in salinity that would affect the $\delta^{11}\text{B}$ record; however,
682 these depositional cycles are not on the same scale as the secular trends seen in $\delta^{11}\text{B}$.

683

684 *3.2.3 Water Depth*

685 Permo-Triassic seawater pH values have been reconstructed assuming carbonate precipitation at
686 sea surface conditions. Modern bathymetric pH and temperature profiles from the tropics show
687 that pH can decrease by up to ~0.2 units and temperature can drop by 15°C in the first 300 m
688 (53). Thus the B isotope composition of carbonates precipitated at greater water depths would be
689 more negative compared to surface precipitates. This potential depth effect is thought to be of
690 minimal influence to the Wadi Bih carbonates due to the exclusively shallow marine (subtidal to
691 intertidal) nature of the succession.

692 Consequently, uncertainties in the $\delta^{11}\text{B}$ -ocean pH relationship potentially introduced by
 693 changes in temperature, salinity, bathymetry or biofacies would not be able to produce the
 694 observed B isotope pattern with the sharp increase in the mid-Changhsingian (~252.04 Ma) and
 695 the temporary decrease in the mid-Griesbachian (~251.88 Ma), but would be capable of
 696 explaining some of the slight variations across the PTB.

697

698 **4. Boron-pH calculations**

699 The boron isotope composition of marine carbonates is used to reconstruct seawater pH values
 700 and atmospheric pCO₂ concentrations (e.g. 54-58). The application of the B isotope system as a
 701 proxy for ocean pH is based on the observation that the fractionation factor for isotope exchange
 702 between the two different speciation of boron (B(OH)₃ and B(OH)₄⁻) is pH sensitive and
 703 produces significant and traceable changes in the isotopic composition of carbonates (e.g. 57).

704 The carbonate-associated boron isotopic composition $\delta^{11}\text{B}_{\text{carb}}$ is assumed to be equal to
 705 that of seawater B(OH)₄⁻. The isotopic composition of the two boron species B(OH)₄⁻ and
 706 B(OH)₃ is determined by the isotopic fractionation factor $\alpha_{\text{B3-B4}} = 1.0272$ (59), the speciation (a
 707 strong function of pH, with smaller dependencies on temperature, pressure and salinity), and the
 708 overall isotopic composition of seawater $\delta^{11}\text{B}_{\text{sw}}$. The relationship between pH and $\delta^{11}\text{B}_{\text{carb}}$ is
 709 given by:

$$710 \quad \text{pH} = \text{p}K_{\text{B}} - \log \left[- \frac{\delta^{11}\text{B}_{\text{sw}} - \delta^{11}\text{B}_{\text{carb}}}{\delta^{11}\text{B}_{\text{sw}} - \alpha_{\text{B3-B4}} \cdot \delta^{11}\text{B}_{\text{carb}} - 1000 \cdot (\alpha_{\text{B3-B4}} - 1)} \right]$$

711 where the effective dissociation constant $\text{p}K_{\text{B}}$ includes the temperature, pressure and salinity
 712 dependence. We dynamically calculate $\text{p}K_{\text{B}}$ using CO2SYS-MATLAB v1.1 (60) (detailed in SM
 713 5), which is consistent with the calculation and constants used by (61).

714 A prerequisite for reconstruction of past ocean pH values via B isotope data is the
715 knowledge of the primary B isotope composition $\delta^{11}\text{B}_{\text{SW}}$ of the ambient seawater. The B isotope
716 composition of modern seawater is regarded as being homogeneous with a $\delta^{11}\text{B}$ value of 39.5‰
717 (54). However, recent work on Precambrian to Phanerozoic carbonates and evaporites has shown
718 that the B isotope composition of seawater was highly variable in the geological past (16, 45, 46)
719 driven mainly by variations in the global boron budget during Earth history (23), exceeding the
720 residence time of B in seawater. Permo-Triassic seawater have been calculated based on changes
721 in processes controlling the oceanic boron budget, e.g. oceanic crust production rate and
722 continental boron flux (16) and suggests $\delta^{11}\text{B}_{\text{SW}}$ values of ~38‰ for the early Permian (285 Ma,
723 Late Sakmarian) and ~34‰ just before the PTB (16).

724 Given this uncertainty in $\delta^{11}\text{B}_{\text{SW}}$, we consider a range of values, determined by the
725 overall consistency of the $\delta^{11}\text{B}_{\text{carb}}$ data with model-derived constraints on the seawater carbonate
726 chemistry and hence pH, resulting in an envelope for the pH calculation. The model scenarios
727 encompass the range of previous estimates for background Early Permian conditions (13), which
728 suggest either high $p\text{CO}_2$ with lower pH (CO_2Hi : ~10 PAL, pH ~7.5), or low $p\text{CO}_2$ with higher
729 pH (CO_2Lo : ~3 PAL, pH ~8). In our model we consider these two scenarios with corresponding
730 seawater boron isotopic composition ($\delta^{11}\text{B}_{\text{SW}} = 34\text{‰}$ and 36.8‰) required to reproduce the
731 observed $\delta^{11}\text{B}_{\text{carb}}$ (see SM Section 6 for further details of the model scenarios).

732 We illustrate the effect of uncertainties in $\delta^{11}\text{B}_{\text{SW}}$ and temperature on pH, and the overall
733 consistency with the $\delta^{11}\text{B}_{\text{carb}}$ data in Figure S3 and Table S1 (note that the full model described
734 in SM 4 also includes a dynamic calculation of temperature). The lowest $\delta^{11}\text{B}_{\text{SW}} = 34\text{‰}$ and
735 highest pH considered (scenario CO_2Lo) is effectively constrained by the model limitation on the
736 highest pH values immediately prior to EP1. The highest $\delta^{11}\text{B}_{\text{SW}} = 36.8\text{‰}$ and lowest pH

737 (scenario CO_2Hi) is effectively constrained by the limit of propagated uncertainty (2σ) on the
738 lowest observed $\delta^{11}B_{carb}$ combined with the non-linearity of the B isotope system.

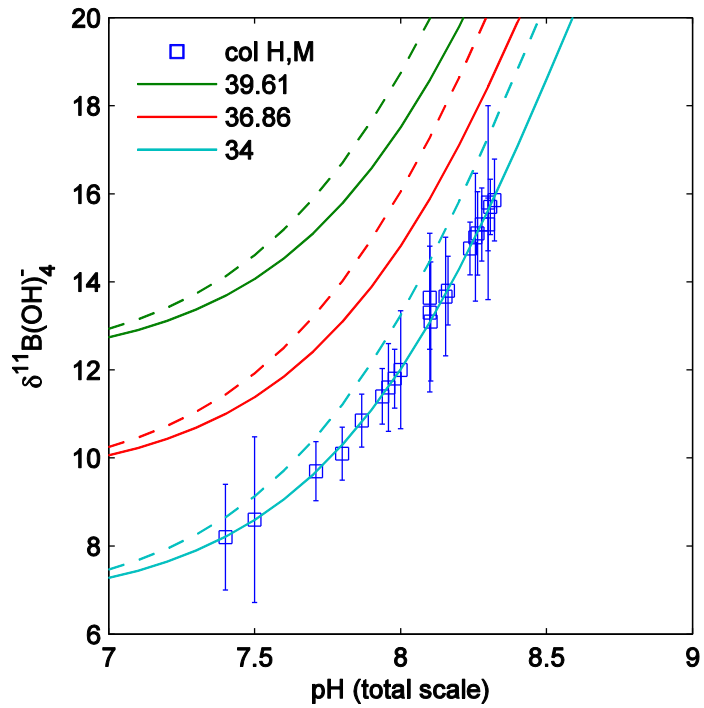
739 An additional uncertainty in interpreting $\delta^{11}B_{carb}$ is introduced by any potential offset or
740 fractionation between seawater borate $\delta^{11}B(OH)_4^-$ and $\delta^{11}B_{carb}$. As shown by (59), the laboratory
741 measured fractionation of abiotic carbonate (61) shows both an offset relative to that of the
742 artificial seawater, and a shallower slope with increasing pH. As we consider $\delta^{11}B_{SW}$ as a model
743 parameter to be determined, an overall offset is effectively subsumed within this. However a
744 shallower slope than that used here ($^{11-10}K_B = 1.0272$ [ref 59], cf the one given for inorganic
745 carbonates (1.0267; [ref 62])) would require a larger pH change for a given observed change in
746 $\delta^{11}B_{carb}$. Sanyal et al. (63) published B isotope values for inorganic calcite precipitates at three
747 different pH conditions - 7.9, 8.3, and 8.6. This same experimental inorganic calcite data and
748 associated estimates of $^{11-10}K_B$ in seawater were also presented by Klochko et al. (59, 64). To
749 date, there is no agreement as to the actual deviation of Sanyal's data (63) from the now
750 generally-accepted empirical fractionation factor given by Klochko et al (59). A best fit value of
751 1.0267 for the inorganic calcite precipitation was given by Pagani et al. (62). But by contrast, the
752 deviation between the inorganic carbonate values and the empirical fractionation appears much
753 bigger in Klochko et al. (59) and the best-fit value is given as 1.0260. This increased deviation is
754 explained by the difference in pH scale between natural or artificial seawater used for the
755 calibration experiments, and the experimental seawater pH measured on the (freshwater) NBS
756 scale. Klochko et al. (64) tries to address the deviations in $\delta^{11}B$ of inorganic precipitates from
757 empirical calibration studies and concludes that boric acid incorporation may contribute to the
758 ^{11}B enrichment observed in inorganic precipitates, especially at lower pH. They point out,

759 however, that all carbonates precipitated under controlled pH conditions were enriched in ¹¹B
760 relative to seawater borate (63).

761 If we use the best fit value (1.0267) given by Pagani et al. (62) for the inorganic
762 carbonates (Sanyal et al 2000) instead of the empirical value of Klochko et al. (59), then the
763 offset for the majority of the data would be around 0.05 pH units. Visible changes to more acidic
764 values would only appear at the lowermost pH estimates, but would still be less than 0.2 pH units
765 and fall within the general uncertainty. If we take the 1.026 value, the offset would generally
766 increase between ≤ 0.1 and 0.2 (pH range between ~ 8.4 and 7.8) but would indeed result in
767 significantly lower pH values of up to 0.5 units for the acidification event.

768 Given the limited amount of experimental inorganic calcite data and this disagreement
769 between the best-fit values, we have taken the simplest approach for the model representation
770 and chosen to use the most generally-accepted empirical fraction factor given by Klochko et al.
771 (59).

772



773

774 **Figure S3** Effect of background $\delta^{11}\text{B}_{\text{SW}}$ and temperature on relationship between $\delta^{11}\text{B}_{\text{carb}}$ and pH
 775 (total scale), for conditions appropriate to low-latitude surface ocean (atmospheric pressure,
 776 salinity 35 psu, temperature 25°C (solid lines) and 35°C (dashed lines)). Three values are shown for
 777 $\delta^{11}\text{B}_{\text{SW}}$: 39.61‰ (the contemporary value) and two values consistent with the end Permian data.
 778 Data points and 2σ errors are overlaid for the $\delta^{11}\text{B}_{\text{SW}}$ 34‰ case.

779

780 **5. Model Description**

781 **5.1 Overview**

782 The overall model structure (shown in Figure S4) is essentially a superset of carbon cycle models
783 previously applied to the end Permian (65, 66, 67), with additional consideration of the marine
784 sulphur cycle. The model includes a three box ocean model as the minimum needed to
785 demonstrate the effect of the biological pump on vertical DIC gradients and ocean redox state.
786 The model is implemented as a set of coupled differential equations for the time evolution of
787 reservoirs (Table S3), exchanging fluxes according to air-sea exchange, ocean circulation,
788 applied external forcings and perturbations, and the biogeochemical processes described in Table
789 S5. Model constants are defined in Table S4. The model implements an open inorganic carbon
790 cycle, with atmospheric $p\text{CO}_2$ and marine DIC and Alk determined by the feedbacks between
791 land-surface carbonate and silicate weathering and marine carbonate deposition. Oxidative
792 weathering, volcanic degassing, and land and marine organic carbon burial are specified as
793 forcings, as is marine phosphorus and hence productivity.

794

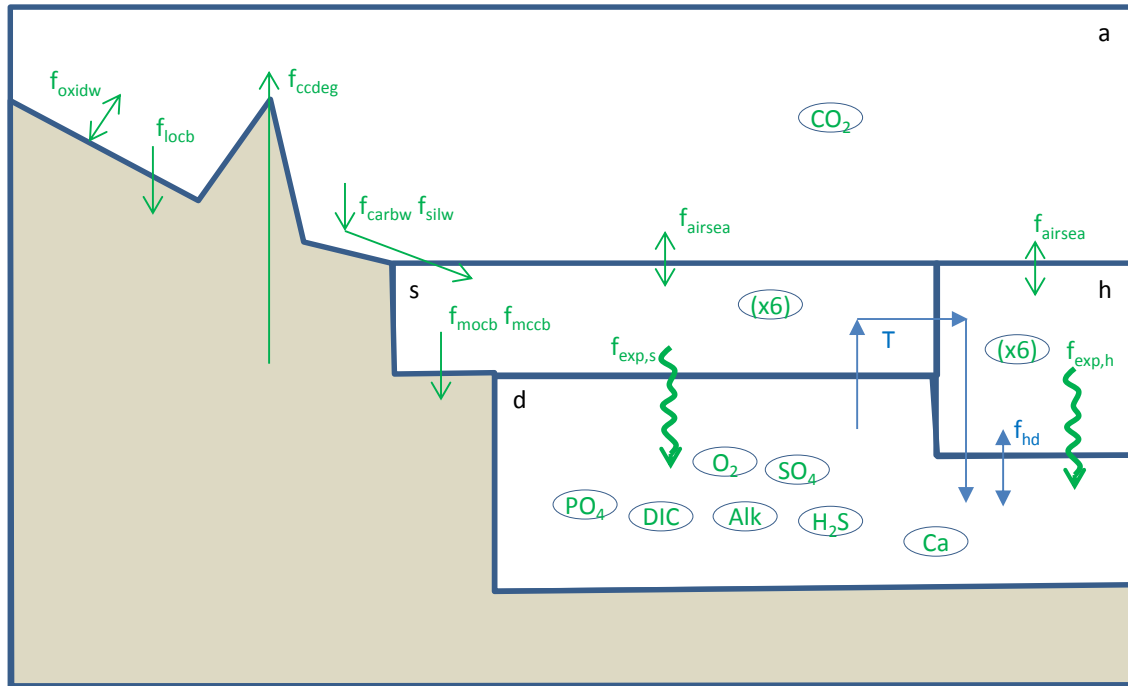


Figure S4 Model schematic. Ocean is represented by three boxes (s, h, d), with well-mixed atmosphere box a. Biogeochemical fluxes are shown in green, ocean circulation in blue. Biogeochemical reservoirs are shown as ovals.

795

796 Table S3 Model reservoirs

Reservoir	Initial size	Description
Atmosphere		
CO ₂	eqb ^a	Atmospheric CO ₂
Ocean		
DIC _(s,h,d)	eqb ^a	Total carbon
Alk _(s,h,d)	eqb ^a	Total alkalinity
O ₂ _(s,h,d)	eqb ^a	Oxygen
PO ₄ _(s,h,d)	2.15 μM / kg-sw	Phosphate
SO ₄ _(s,h,d)	28 mM / kg-sw	Sulphate
H ₂ S _(s,h,d)	0	Sulphide
Ca	10.28 mM / kg-sw	Calcium (single reservoir)

^a set by initial spin-up to steady-state.

797

798 Table S4 Model fluxes, constants, and forcings.

Flux	Model Parameter	Value	Units	Description
crust/mantle				
f_{ccdeg}	k_{12_ccdeg}	varies ^a	mol CO ₂ yr ⁻¹	Degassing
land surface				
f_{carbw}	k_{14_carbw}	1.33x10 ¹³	mol CaCO ₃ yr ⁻¹	Carbonate weatherability
f_{silw}	k_{silw}	varies ^a	mol "CaSiO ₃ " yr ⁻¹	Silicate weatherability
f_{oxidw}	k_{17_oxidw}	varies ^a	mol C yr ⁻¹	Oxidative weathering
f_{locb}	k_{5_locb}	5x10 ¹²	mol C yr ⁻¹	Land organic carbon burial
marine				
	$k_{O_Asurf(s)}$ $k_{O_Asurf(h)}$	0.85x3.6 x10 ¹⁴ 0.15x3.6 x10 ¹⁴	m ²	Ocean surface area
	$k_{O_vol(s)}$ $k_{O_vol(h)}$ $k_{O_vol(d)}$	3.06 x10 ¹⁶ 1.35 x10 ¹⁶ 1.60 x10 ²¹	m ³	Ocean volume
T	k_{O_circT}	20	Sv	Overturning circulation
f_{hd}	$k_{O_circfhd}$	60	Sv	High latitude – deep exchange
	k_{piston}	0.3	m d ⁻¹	Air-sea piston velocity
	$T_{O(s)}$ $T_{O(d)}$	$T_{\text{mean}} + 6.5$ $\max(T_{\text{mean}} - 12.5, 2.5)$		Ocean temperature
	k_{O_sal}	35	psu	Ocean salinity
	$r_{c:p}$	161		Redfield C:P ratio
	$r_{n:p}$	16		Redfield N:P ratio
f_{mccb}	$k_{\text{carbsedshallow}}$	varies ^a	mol CaCO ₃ yr ⁻¹	Marine carbonate burial
f_{moch}	k_{2_moch}	5x10 ¹²	mol C yr ⁻¹	Marine organic carbon burial

^a parameters k_{silw} , k_{12_ccdeg} , $k_{\text{carbsedshallow}}$ are per-scenario external forcings.

799

800

Table S5 Biogeochemical processes

Flux	Biogeochemical transformation	Stoichiometry							Rate	Description	
		CO _{2(a)}	DIC	Ca	O ₂	P	H ₂ S	SO ₄ ²⁻			Alk
<i>land surface</i>											
f _{carbw}	CO ₂ + H ₂ O + CaCO ₃ → Ca ²⁺ + 2HCO ₃ ⁻	-1	2	1					2	Equation (S2)	Carbonate weathering
f _{silw}	2CO ₂ + H ₂ O + CaSiO ₃ → SiO ₂ + Ca ²⁺ + 2HCO ₃ ⁻	-2	2	1					2	Equation (S1)	Silicate weathering
f _{oxidw}	CH ₂ O + O ₂ → CO ₂ + H ₂ O	1	0	0					0	k _{17_oxidw}	Oxidative weathering
f _{locb}	CO ₂ + H ₂ O → CH ₂ O + O ₂	-1	0	0					0	k _{5_locb}	Organic carbon burial
<i>marine</i>											
f _{exp}	CO ₂ + H ₂ O + xP → CH ₂ OP _x + O ₂		-1	0	1+	-1/r _{cp}	0	0	r _{np} /r _{cp} ^a	see Section 3.2.4	Export production
	CH ₂ OP _x + O ₂ → CO ₂ + H ₂ O + xP		1	0	2r _{np} /r _{cp}	1/r _{cp}	0	0	-r _{np} /r _{cp} ^a	see Section 3.2.4	Aerobic remineralisation
	CH ₂ OP _x + $\frac{1}{2}$ SO ₄ ²⁻ → HCO ₃ ⁻ + xP + $\frac{1}{2}$ H ₂ S		1	0	2r _{np} /r _{cp}	1/r _{cp}	0.5	-0.5	1	see Section 3.2.4	Sulphate reduction
	H ₂ S + 2O ₂ → SO ₄ ²⁻ + 2H ⁺		0	0	-2	0	-1	1	-2	see Section 3.2.4	Sulphide oxidation
f _{mccb}	Ca ²⁺ + HCO ₃ ⁻ → H ⁺ + CaCO ₃	-1	-1	0	0	0	0	0	-2	k _{carbseeshallow} x(Ω _{aragonite} -1) ^{1.7}	Carbonate deposition
	2H ₂ S + Fe(OH) ₃ → FeS ₂ + $\frac{1}{2}$ H ₂ + 3H ₂ O	0	0	0	-0.5	0	-2	0	0		Pyrite formation and burial
f _{mocb}	CO ₂ + H ₂ O → CH ₂ O + O ₂	-1	0	1					0	k _{2_mocb}	Organic carbon burial

^a contribution from nitrate, assumed at Redfield ratio.

591 **5.2 Detailed description**

592 *5.2.1 pCO₂ and temperature*

593 Global mean temperature T_{mean} is calculated from atmospheric pCO₂ using the energy
594 balance model as in COPSE (68, 69), with solar insolation appropriate for 250 Mya. Marine
595 temperature $T_{\text{O(s,h,d)}}$ is calculated from the global mean assuming a fixed high-low latitude
596 temperature difference and minimum temperature of 2.5⁰C, with $T_{\text{O(s)}}=T_{\text{mean}}+6.5$ ⁰C and
597 $T_{\text{O(h,d)}}=\max(T_{\text{mean}}-12.5,2.5)$ ⁰C.

598

599 *5.2.2 Land surface weathering, degassing, and organic carbon burial*

600 Land surface weathering is essentially a stripped-down version of that in COPSE (69), with
601 carbonate and silicate weathering included but sulphur weathering and the long-timescale
602 sedimentary reservoirs for sulphur and carbon are omitted. Atmospheric oxygen is fixed at
603 the present-day value. Silicate and carbonate weathering are controlled by atmospheric pCO₂
604 and temperature with functional forms:

$$f_{\text{silw}}=k_{\text{silw}} e^{0.090(T_{\text{mean}}-T_0)} [1 + 0.038(T_{\text{mean}} - T_0)]^{0.65} \left(\frac{2 \text{ pCO}_2}{1 + \text{ pCO}_2} \right) \quad (\text{S1})$$

$$f_{\text{carbw}}=k_{14_carbw} [1 + 0.087(T_{\text{mean}} - T_0)] \left(\frac{2 \text{ pCO}_2}{1 + \text{ pCO}_2} \right) \quad (\text{S2})$$

605 where $T_0=15$ ⁰C.

606

607 Atmospheric CO₂ is consumed, and carbon, alkalinity and calcium supplied to the ocean,
608 according to the stoichiometry defined in Table S5. Volcanic degassing, oxidative
609 weathering, and land organic carbon burial are prescribed as forcings.

610

611

612

613 5.2.3 Marine circulation and air-sea exchange

614 Marine circulation is defined by a thermohaline circulation T , and high-latitude – deep
615 exchange f_{hd} . The marine box model includes P , O , DIC , Alk , SO_4 , H_2S as per-box prognostic
616 variables, along with DIC $\delta^{13}C$. Marine carbonate chemistry is calculated using a modified
617 version of CO2SYS-MATLAB v1.1 (60), extended to include sulphide alkalinity using the
618 constants tabulated by (70). Air-sea exchange of oxygen and CO_2 assume a fixed piston
619 velocity and temperature-dependent solubility.

620

621 5.2.4 Marine productivity and burial

622 Productivity and hence the biological pump are controlled by the marine phosphorus
623 circulation. Biomass is produced in the surface ocean boxes, and exported and remineralized
624 in the deep box according to the processes and stoichiometries defined in Table S5.
625 Productivity in the ‘s’ box is specified to consume all phosphorus down to negligible
626 concentration. Productivity in the ‘h’ box is specified to consume a fraction 0.18 of input
627 phosphorus. Remineralization in the deep ocean box consumes oxygen down to a limiting
628 concentration of $1\mu M$ and thereafter reduces sulphate to sulphide.

629 Shallow-water carbonate deposition occurs in the ‘s’ box only and is controlled by
630 aragonite saturation state and a parameter representing a combination of shelf area and
631 calcification effectiveness, ie a ‘Neritic’ ocean without pelagic calcifiers (71), functional
632 form $k_{carbsedshallow} * (\Omega_{aragonite} - 1)^{1.7}$. Ocean carbonate compensation is not included, on the basis
633 this will be small for a ‘Neritan’ ocean.

634 Marine organic carbon burial from the ‘s’ box and pyrite burial from the ‘d’ box are
635 imposed as forcings. Given the uncertainties in controls on phosphorus input over the end-
636 Permian interval, the marine phosphorus cycle is semi-closed (i.e. weathering feedback on
637 phosphorus input is not included), with imposed perturbations (forcings) to source/sink

638 balance (riverine input vs burial) applied as inputs to the ‘s’ box to control phosphorus
639 concentration.

640

641 5.2.5 Carbon isotopes

642 Carbon isotopes are implemented with additional reservoirs for each of atmospheric CO₂ and
643 marine DIC_(s,h,d). Air-sea exchange fractionates according to the temperature-dependent
644 equilibrium and kinetic factors determined by Zhang et al. (72). Marine export production is
645 at fixed fractionation relative to DIC, with $\delta^{13}\text{C}_{\text{exp}(s,h)} = \delta^{13}\text{DIC}_{(s,h)} - 25\text{‰}$.

646 Volcanic degassing, carbonate weathering, and oxidative weathering are assumed to
647 add carbon at fixed fractionations of -4.9‰, 2.65‰, and -25‰ respectively. Atmospheric
648 CO₂ consumed by silicate and carbonate weathering (and added to the ocean ‘s’ box as DIC)
649 is fractionated relative to atmospheric CO₂ according to the freshwater fractionation of Zhang
650 et al. (72).

651 Land and marine organic carbon burial is at fixed fractionation relative to atmospheric
652 CO₂ and marine DIC, with $\delta^{13}\text{C}_{\text{locb}} = \delta^{13}\text{CO}_2 - 19\text{‰}$ and $\delta^{13}\text{C}_{\text{moch}} = \delta^{13}\text{DIC}_{(s)} - 25\text{‰}$. Marine
653 inorganic carbon burial is assumed to not fractionate relative to DIC_(s).

654

655 5.2.6 Boron isotopes

656 Carbonate-associated boron isotopic composition $\delta^{11}\text{B}_{\text{carb}}$ is assumed to be equal to that of
657 seawater B(OH)₄⁻. Speciation of B(OH)₄⁻ and B(OH)₃ is calculated using CO2SYS-MATLAB
658 v1.1 (60). Isotopic composition is then calculated from speciation $x_{\text{bo4}} = [\text{B}(\text{OH})_4^-]/[\text{B total}]$
659 and seawater $\delta^{11}\text{B}_{\text{sw}}$ as

$$\delta^{11}\text{B}_{\text{carb}} = \frac{\delta^{11}\text{B}_{\text{sw}} - 1000(1 - x_{\text{bo4}})(\alpha_B - 1)}{\alpha_B - x_{\text{bo4}}(\alpha_B - 1)}$$

660 with the isotopic fractionation factor $\alpha_B = 1.0272$.

661

662

663 *5.2.7 Model spinup and steady state*

664 The model pCO₂ steady-state is defined by the imposed degassing rate, organic carbon burial
665 and oxidation, and silicate weathering parameterisations, where these together define a
666 unique value for the steady-state atmospheric pCO₂ and temperature where $f_{\text{ccdeg}}+f_{\text{oxidw}} =$
667 $f_{\text{silw}}+f_{\text{locb}}+f_{\text{mocb}}$. During spin-up, the ocean chemistry and hence atmosphere-ocean CO₂
668 partitioning adjust (on the silicate weathering timescale of ~100kyr) to a steady-state, with
669 the aragonite saturation state adjusting such that carbonate burial balances inputs from
670 carbonate and silicate weathering.

671

672 *5.2.8 Sensitivity to seawater composition*

673 The Mg and Ca composition of end-Permian seawater (as determined from fluid inclusions)
674 is consistent with that of modern seawater ([Mg] = 53 mmol/kg, [Ca] = 10.3 mmol/kg), but
675 with large uncertainties (6). Sulphate concentration may have been much lower than modern
676 values (73, 74). We estimate the uncertainties in carbonate system chemistry following the
677 approach of (75). We estimate the effect on K_1 and K_2 from model results (74) as:

$$\frac{K_1^*}{K_1} = 1 + 0.155 \frac{\Delta[\text{Mg}^{2+}]}{[\text{Mg}^{2+}]_m} + 0.033 \frac{\Delta[\text{Ca}^{2+}]}{[\text{Ca}^{2+}]_m} - 0.019 \frac{\Delta[\text{SO}_4^{2-}]}{[\text{SO}_4^{2-}]_m}$$

$$\frac{K_2^*}{K_2} = 1 + 0.641 \frac{\Delta[\text{Mg}^{2+}]}{[\text{Mg}^{2+}]_m} + 0.071 \frac{\Delta[\text{Ca}^{2+}]}{[\text{Ca}^{2+}]_m} - 0.054 \frac{\Delta[\text{SO}_4^{2-}]}{[\text{SO}_4^{2-}]_m}$$

678 where subscript m refers to modern values and an asterix to end-Permian values. Taking
679 extremal values $[\text{Mg}^{2+}]^*/[\text{Mg}]_m=0.5$, $[\text{Ca}^{2+}]^*/[\text{Ca}]_m=2$, $[\text{SO}_4^{2-}]^*/[\text{SO}_4^{2-}]_m=0.1$, we have
680 $K_1^*/K_1=1-0.078+0.033+0.0171=0.972$ or $\text{p}K_1^*=\text{p}K_1+0.012$ (or for $[\text{Mg}^{2+}]$ alone,
681 $K_1^*/K_1=0.92$, $\text{p}K_1^*=\text{p}K_1+0.035$). For a fixed pCO₂, this implies a corresponding decrease in
682 $[\text{HCO}_3^-]$ and hence DIC inventory, or equivalently an increase in pH for the same DIC
683 content. The corresponding effect on K_2 is much larger, $K_2^*/K_2=1-$

684 0.32+0.0171+0.049=0.75. We estimate the effect of Mg concentration on the calcite solubility
685 constant K_{sp} using the parameterisation given by (76) as:

$$\frac{K_{sp}^*}{K_{sp,m}} = 1 - 0.0833 \left(\frac{[Mg^{2+}]_m}{[Ca^{2+}]_m} - \frac{[Mg^{2+}]^*}{[Ca^{2+}]^*} \right)$$

686 which gives $K_{sp}^*/K_{sp}=1-0.33$.

687 This sensitivity study shows that the uncertainty in K_1 therefore introduces only a small
688 uncertainty in pH ($\sim < 0.03$) or equivalently a $\sim 10\%$ uncertainty in DIC inventory relative to
689 pCO_2 . The combined uncertainties in K_2 , K_{sp} and $[Ca^{2+}]$ are larger, introducing corresponding
690 uncertainties in calcite saturation state. However, within the model employed here, this is
691 effectively absorbed into a rescaling of the parameter $k_{carbsedshallow}$.

692

693 **6- Model Scenarios**

694 Given the uncertainty in the absolute value of the seawater boron isotope composition
695 (section S4), we consider two scenarios with initial steady-state conditions as in Table S6:
696 scenario *CO₂Hi* uses $\delta^{11}B_{SW}=36.86$, and *CO₂Lo* uses $\delta^{11}B_{SW}=34$. Here ocean pH depends on
697 the combination of atmospheric pCO_2 and carbonate saturation state. Atmospheric pCO_2 (set
698 by the ratio of net carbon sources : silicate weatherability) is tuned to 3 or 10 PAL,
699 encompassing the range of conditions considered by previous model studies (77, 78, 79).
700 Ocean carbonate saturation state is set by carbonate sedimentation efficiency parameter
701 $k_{carbsedshallow}$, which represents a combination of shelf area x deposition rate. A summary of
702 scenarios required to explain the three events in our $\delta^{11}B$ data are given in Table S7 and
703 explored in further detail below.

704

705 Table S6 Steady-state initial conditions summary

	CO₂Hi	CO₂Lo	Units	Description
<i>Parameters</i>				
k _{12_ccdeg}	11.80x10 ¹²	11.80x10 ¹²	mol CO ₂ yr ⁻¹	degassing
k _{17_oxidw}	5x10 ¹²	5.92x10 ¹²	mol C yr ⁻¹	oxidative weathering
k _{silw}	2.40x10 ¹²	6.60x10 ¹²	mol CaSiO ₃ yr ⁻¹	Silicate weatherability
k _{carbsedshallow}	18.43x10 ¹²	1.44x10 ¹²	mol CaCO ₃ yr ⁻¹	Marine carbonate burial efficiency
<i>Steady-state conditions</i>				
pCO ₂ , ppm	2800	845		
T _{mean}	22.18	14.94	°C	Global mean temp
T _{O(s)}	28.7	21.44	°C	Ocean temp
Ω _{arag(s)}	2.44	6.15		Aragonite saturation, 's' box
pH	7.51,7.46,7.07	8.02, 7.99, 7.60	Total scale	Ocean pH, 's', 'h', 'd'
DIC total	5.53x10 ¹⁸	6.01x10 ¹⁸	mol	
δ ¹³ DIC _(s)	+3.32	+2.97	‰	
<i>Fluxes</i>				
f _{carbw}	27.46x10 ¹²	15.58x10 ¹²	mol CaCO ₃ yr ⁻¹	Carbonate weathering
f _{silw}	6.80x10 ¹²	7.72x10 ¹²	mol "CaSiO ₃ " yr ⁻¹	Silicate weathering
f _{ccdeg}	11.80x10 ¹²	11.80x10 ¹²	mol CO ₂ yr ⁻¹	Degassing
f _{mccb}	34.26 x10 ¹²	23.30x10 ¹²	mol CaCO ₃ yr ⁻¹	Marine carbonate burial

706

707 Table S7 Overview of scenarios

Event	Time	Scenario	
		<i>CO₂Hi</i>	<i>CO₂Lo</i>
		$\delta^{11}\text{B}_{\text{sw}}=36.86$	$\delta^{11}\text{B}_{\text{sw}}=34$
pH rise	252.25 Ma	Marine productivity increase	
	252.05 Ma	Carbonate sedimentation efficiency decrease	
		Pyrite deposition, Carbonate weathering increase	
$\delta^{13}\text{C}$ drop, stable pH	252.00 Ma	Land organic carbon burial decrease	
	251.95 Ma	Slow (~50kyr) isotopically light carbon addition.	
Acidification, $\delta^{13}\text{C}$ change	no 251.88 Ma	Rapid (10kyr) isotopically heavy carbon addition	

708

709

710 **6.1 Mechanisms for pH rise**

711 The rise in $\delta^{11}\text{B}$ at 252.05 Ma corresponds to a rise in pH from 7.5 to 8.0 for scenario *CO₂Hi*,
712 and from 8.0 to 8.3 for scenario *CO₂Lo*. Given the major upheavals in the Earth system at the
713 end-Permian, multiple mechanisms including changes in silicate and carbonate weathering,
714 and marine changes in calcification and due to anoxia may potentially contribute to pH rise,
715 and we use model sensitivity studies to determine potential constraints. We first review
716 evidence for each mechanism, then summarize the model sensitivity studies and propose
717 plausible scenarios involving multiple drivers.

718 *6.1.1 Carbonate weathering*

719 The overall high weathering rates across the Permian-Triassic boundary also imply an
720 increase in carbonate weathering, possibly combined with any direct effect from chemical
721 weathering of eroded clastic material itself.

722 There is also direct evidence for later sea-level regression and subaerial weathering of
723 the shallowest sections of the S. China carbonate platform, coincident with EP1 (20) which
724 may contribute to maintaining pH over this interval. To estimate an order of magnitude for
725 enhanced carbonate shelf weathering: 10^{18} mol CaCO_3 , density 2.8 g cm^{-3} has volume 36,000
726 km^3 , or $100\text{m} \times 3.6 \times 10^5 \text{km}^2$, or 1800km of 200km wide shelf; this would provide a
727 weathering flux of 10^{13} mol $\text{CaCO}_3 \text{ yr}^{-1}$ for 100 kyr, comparable to the background carbonate
728 weathering rate.

729 *6.1.2 Decrease in calcification effectiveness*

730 Overall global biotic calcification effectiveness is determined by the combination of available
731 shelf area, and local ecosystem-dependent rates. A reduction in area of deposition could be
732 because anoxic/euxinic waters extend onto the shelf bottom precluding those parts from
733 hosting deposition, and/or because there is a sea level change reducing shelf area. The input
734 of clastics prior to EP1 could also reduce carbonate production as a result of ecosystem
735 impacts (80, 81).

736 *6.1.3 Productivity-driven ocean anoxia, sulphate reduction, and pyrite burial*

737 Multiple lines of evidence suggest a (large) expansion of oxygen minimum zones prior to the
738 PT boundary, while the deep ocean remains suboxic. These include U isotope evidence for
739 ~6x increase in anoxic fraction (82), extensive pyrite deposition (83), and GCM studies (84)
740 illustrating the spatial distribution of anoxia.

741 We represent productivity-driven ocean anoxia by increasing marine phosphorus from
742 the present-day value to 2.3x present (the value used by (66)). This is achieved by adding
743 phosphorus to the marine 's' reservoir (representing a net excess of riverine input over
744 sediment output) at rate $3.9 \times 10^{10} \text{ mol yr}^{-1}$ over the interval 252.15 – 252.05 Ma (cf present-
745 day riverine input $\approx 2 \times 10^{10} \text{ mol yr}^{-1}$) (64). This results in pCO_2 drawdown as a result of the
746 increased biological pump, and in sulphate reduction leading to $\sim 50 \text{ }\mu\text{M H}_2\text{S}$ in the 'd' box.
747 Note that sulphate reduction increases alkalinity / pH in the 'd' box by $\Delta\text{pH} \sim 0.1$, but as the
748 S redox shuttle is completed by sulphide oxidation at the base of the oxic surface box, this
749 has no effect on the pH of the oxic surface ocean.

750 High rates of pyrite formation are seen at and before EP1 (83, 85). Pyrite burial
751 results in a net alkalinity source (86). We assume that pyrite deposition is a water-column
752 process, hence is limited by the availability of iron and sulphide. The most limiting factor for
753 the scenarios considered here is iron. To quantify this, we estimate pyrite deposition rate
754 sustainable over a timescale of $\sim 100 \text{ kyr}$ as $\sim 1.25 \times 10^{12} \text{ mol FeS}_2 \text{ yr}^{-1}$, based on availability of
755 reactive iron Fe_{HR} (in the contemporary oxic ocean, Fe_{HR} total input to the ocean is $\sim 6.8 \times 10^{12}$
756 mol yr^{-1} , of which only $\sim 1.3 \times 10^{12} \text{ mol yr}^{-1}$ gets to the deep ocean (87). Sulphide availability
757 is determined by anaerobic organic carbon remineralisation in the 'd' box, which is a model-
758 determined fraction of total export production $\sim 10^{14} \text{ mol yr}^{-1}$, and is less limiting than iron for
759 the scenarios considered here. This assumes that sulphate for remineralisation (and hence
760 sulphide availability) is unconstrained by marine sulphur availability, ie that the marine
761 sulphate reservoir is drawn down. This is supported by (or at least consistent with) the data
762 indicating low early Triassic marine sulphate, $< 4 \text{ mM}$ (75, 88).

763

764 *6.1.4 Silicate weathering*

765 Anomalously high sediment fluxes across the Permian-Triassic boundary (80, 81) imply soil
766 loss and increased exposure of highly weatherable rock surfaces, with increases in both
767 physical and chemical weathering. Early Siberian traps emplacement and an increase in area
768 of weatherable basalt also may contribute to an increase in overall silicate weatherability. The
769 main increase in silicate weathering is seen in the Griesbachian but increases may have
770 accompanied the carbon injection for EP1.

771 *6.1.5 Combined scenarios*

772 We summarize the effect of illustrative perturbations for the separate contribution of each
773 potential driver of pH increase in Table S8. Perturbations were applied to the CO₂Lo steady
774 state at 251.95Ma (corresponding to the pH rise seen in the δ¹¹B data), with effect shown
775 100kyr later (at EP1). The marine carbonate system responds on a timescale of ~10kyr,
776 hence reaches a steady state source-sink balance, however the timescale for land-surface
777 weathering feedbacks is >~100 kyr hence the system does not reach a steady state. The
778 dynamic response is included in the full scenarios.

779 As shown in figure S5, the perturbations fall into two groups. Increases in silicate
780 weatherability and pyrite deposition leave carbonate input rate and calcification output
781 essentially unchanged, hence result in changes to atmospheric pCO₂ at nearly constant
782 saturation state. Changes to atmospheric pCO₂ alone are unable to produce a change in pH or
783 δ¹¹B as large as that seen in the data without implausibly low pCO₂ and hence temperature,
784 hence these cannot be the sole drivers of pH increase. Reductions in calcification
785 effectiveness or increases in carbonate inputs change both the carbonate system saturation
786 state and atmospheric pCO₂, resulting in a much larger increase in pH for a given decrease in
787 pCO₂. A major contribution from either or both of these mechanisms is therefore required.

788 The results from the perturbations study demonstrate that all mechanisms may
789 contribute to pH rise, with a major contribution from either or both carbonate input and
790 decrease in calcification effectiveness required. Given the most likely case where multiple
791 mechanisms all contribute, we construct scenarios with a decrease in calcification
792 effectiveness as the major component and smaller contributions from pyrite burial and
793 carbonate weathering (Figure S6). The perturbations considered for the two scenarios are
794 summarized in Table S9 and the dynamic response is shown in Figure S6.

795

796

797

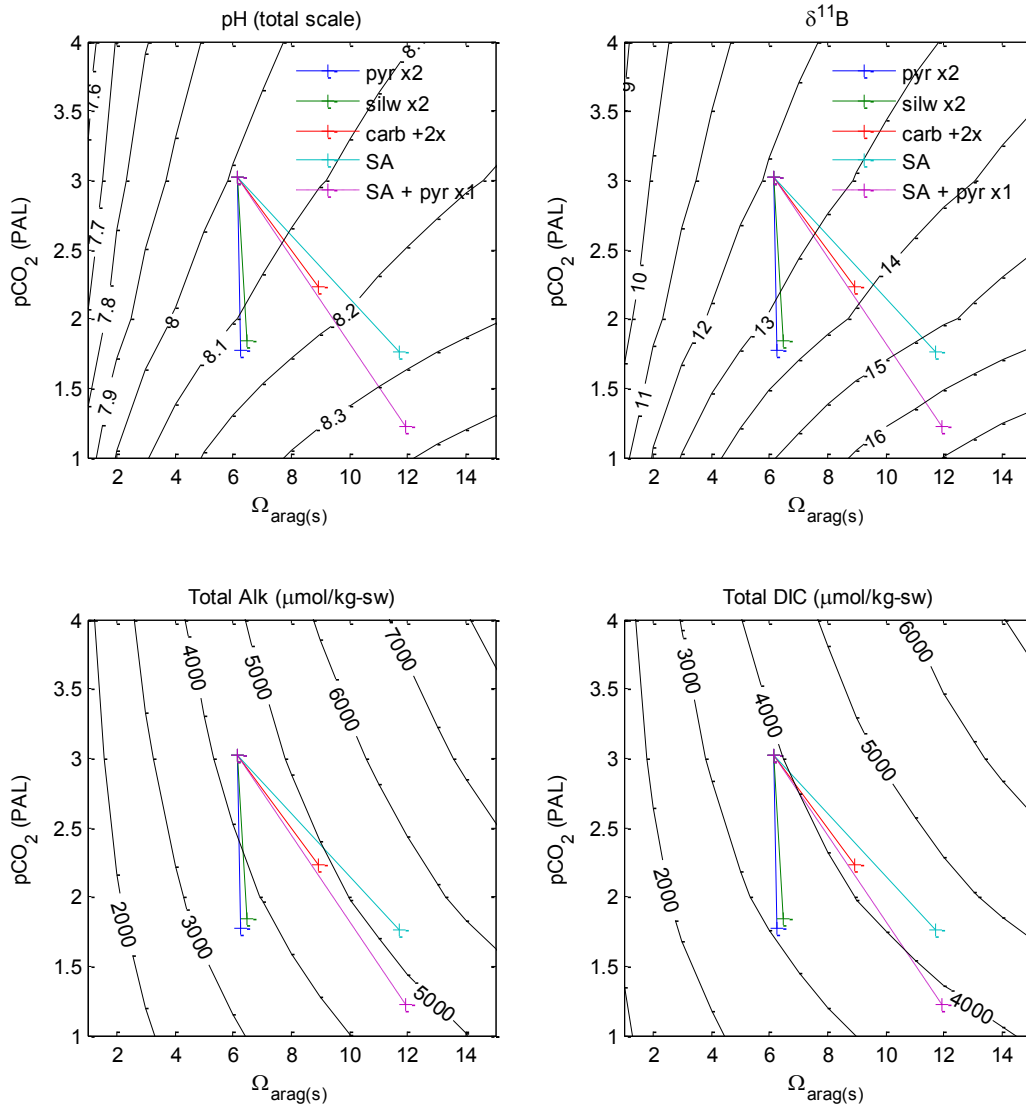


Figure S5 Effect of illustrative perturbations (defined in Table S8) on carbonate system properties for atmosphere and ocean surface (s) box. Contours show carbonate system parameters for a constant temperature of 25 °C, salinity 35 psu, pressure 1 atm, hence do not exactly correspond to the full model results with varying temperature.

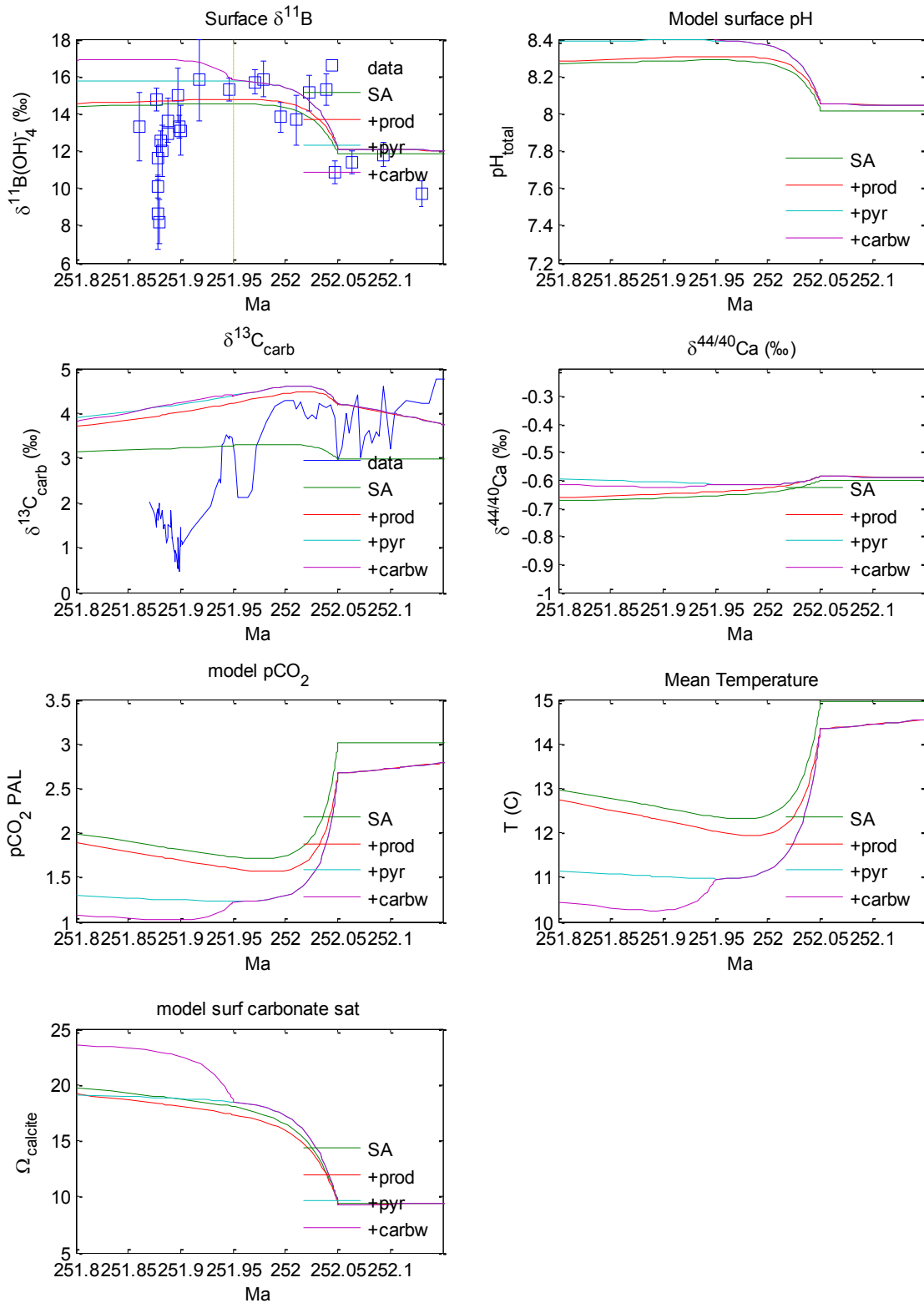


Figure S6 Scenario CO_2Lo : Contributions to pH increase. 'SA': effect of calcification effectiveness decrease. '+prod' effect of increased marine productivity and anoxia. '+pyr' pyrite burial, and '+carb' carbonate weathering

Table S8 Sensitivity study for pH rise mechanisms. Perturbations applied to steady-state CO₂Lo at 252.05Ma, effect shown at 251.95Ma.

Mechanism	Perturbation	pH_{tot(s)}	δ¹¹B(s)	Ω_{arag(s)}	pCO₂	T_{mean}	Comments
<i>steady-state</i>	none	8.02	11.83	6.15	845	14.94	Error! Reference source not found.
CO₂Lo		Δ	Δ	Δ	Δ	Δ	
'calc eff'	K _{carbsedshallow} 14.4x10 ¹¹	→ +0.266	+2.72	+5.55	-351	-2.58	
decrease	2.47x10 ¹¹ (ie x 0.17)						
	K _{carbsedshallow} 14.4x10 ¹¹	→ +0.36	+3.78	+8.2	-446	-3.39	
	1.23x10 ¹¹ (ie x 0.085)						
pyr deposition	'x1': 1.25x10 ¹² mol FeS ₂ yr ⁻¹	+0.083	+0.71	-0.03	-213	-1.39	
	'x2': 2.5x10 ¹² mol FeS ₂ yr ⁻¹	+0.155	+1.36	+0.12	-347	-2.46	
	'x4': 5x10 ¹² mol FeS ₂ yr ⁻¹	+0.318	+3.062	+0.46	-557	-4.66	~ upper limit from Fe availability
carb input	'+1x': add 15x10 ¹² mol CaCO ₃ yr ⁻¹	+0.082	+0.77	+1.44	-130	-0.81	+1x = approx 2x background weathering rate. Ocean carb deposition rapidly compensates
	'+2x': add 30x10 ¹² mol CaCO ₃ yr ⁻¹	+0.147	+1.41	+2.76	-219	-1.44	+2x = ~3x background weath rate
	'+3x': add 45x10 ¹² mol CaCO ₃ yr ⁻¹	+0.20	+1.97	+3.99	-285	-1.94	+3x = ~4x background weath rate
silw increase	x2 k _{silw} 6.60x10 ¹² → 13.2x10 ¹²	+0.14	+1.19	+0.35	-329	-2.30	'Weatherability': drives pH primarily via pCO ₂ decrease
calc eff + pyr	x4 k _{silw} 6.60x10 ¹² → 26.4x10 ¹²	+0.32	+3.08	+0.57	-588	-5.09	
	K _{carbsedshallow} 14.4x10 ¹¹	→ +0.37	+3.91	+5.78	-501	-3.98	used for Scenario CO ₂ Lo.
	2.47x10 ¹¹						
carb + pyr	'x1': 1.25x10 ¹² mol FeS ₂ yr ⁻¹	+0.28	+3.09	+2.92	-485	-3.79	
	'+2x': add 30x10 ¹² mol CaCO ₃ yr ⁻¹						
	'x2': 2.5x10 ¹² mol FeS ₂ yr ⁻¹						

804

805

806 Table S9 Summary of pH rise mechanisms included in scenarios *CO₂Hi* and *CO₂Lo*

	<i>CO₂Hi</i>	<i>CO₂Lo</i>
Decrease in calcification effectiveness	$k_{\text{carbsedshallow}}$ decrease from 1.8433×10^{13} to 0.72×10^{12}	$k_{\text{carbsedshallow}}$ decrease from 1.44×10^{12} to 2.47×10^{11}
Productivity, anoxia	Increase P linearly to 2.3x, over interval 252.25 – 252.05 Ma	
Pyrite burial	1.25×10^{12} mol FeS ₂ yr ⁻¹ starting at 252.05Ma	
Carbonate weathering	9.7×10^{12} mol/yr starting at 251.95Ma	

807

808

809

810 ***6.2 Sources of carbon and volatiles causing a negative $\delta^{13}\text{C}$ excursion or pH excursion***

811

812 ***6.2.1 Terrestrial Carbon burial***

813 The effect of destruction of the land biota and a corresponding decline in land organic
814 carbon burial at the PT boundary was first considered by Broecker and Peacock (89). We
815 represent this by setting land organic carbon burial to zero at $T=251.95\text{Ma}$, resulting in a
816 drop in $\delta^{13}\text{C}_{\text{carb}}$ of $\approx 4\text{‰}$ (comparable to that seen in end-Permian sections), but over
817 timescale of $\sim 100\text{kyr}$.

818

819 ***6.2.2 Marine Carbon burial***

820 Despite (or because of) ecological shifts to a cyanobacteria dominated ecosystem (3), proxy
821 evidence for continued high sediment organic carbon fluxes (84) and a vertical $\delta^{13}\text{C}$
822 gradient (92, 93) as well as continuing marine anoxia suggests that high marine
823 productivity continues through Permian-Triassic. Interruption to the marine biological
824 pump would result in a short timescale (ocean circulation timescale $\sim 10^3$ yr) increase in
825 surface ocean pH and decrease in surface ocean (and hence carbonate) $\delta^{13}\text{C}$. This was
826 considered as a mechanism for short-timescale $\delta^{13}\text{C}$ fluctuations over the PT interval (67).
827 We do not consider this further here as this mechanism results in coupled perturbations to
828 both surface ocean pH and $\delta^{13}\text{C}$, and hence cannot be the major driver for the ‘ $\delta^{13}\text{C}$ without
829 pH’ or ‘pH without $\delta^{13}\text{C}$ ’ signals. Smaller perturbations (interruptions) to the biological
830 pump may be plausible as a contributor to short-timescale $\delta^{13}\text{C}$ signals following EP1.

831

832 *6.2.3 Siberian traps volcanism and contact metamorphism*

833 In order to bound plausible model scenarios, we review here potential mechanisms for
834 volatile input from the Siberian Traps and summarise constraints on magnitudes and rates,
835 based on (24, 27, 93).

836 The Siberian Traps magma area is estimated as $2.5 - 5 \times 10^6 \text{ km}^2$ and volume $>$
837 $2 \times 10^6 \text{ km}^3$. This intruded into the Tunguska sedimentary sequence, which reaches 12.5 km in
838 thickness and includes. ≈ 2.5 km of Cambrian evaporites containing abundant limestone,
839 halite, dolomite and anhydrite (24, 27), coal deposits, as well as Neo-Proterozoic
840 petroleum-bearing shale and carbonate. Intrusion of magma into the sediments resulted in
841 abundant sills and dykes with accompanying contact aureoles, and explosively-generated
842 pipes. The igneous province is estimated to contain approximately 50% intrusive dykes and
843 sills, 30% basalt lava flows, and 20% pyroclastic material (94, 95).

844 The volume of sediments affected by contact metamorphism is estimated (27) from
845 a sill area of $2 \times 10^6 \text{ km}^2$, thickness 200m, generating a contact aureole of thickness 400m.
846 This results in potential release of $0.8 - 2.3 \times 10^{18}$ mol C (as isotopically light methane and
847 CO_2) from metamorphic degassing of organic carbon (assuming the TOC weight percent
848 reacted is 0.5 – 1.5%), over a timescale ≈ 50 kyr. In addition, pipes (each with source region
849 $\approx 5 \text{ km}^3$) could release $0.1 - 0.3 \times 10^{18}$ mol C over ≈ 6.5 kyr. Intrusion into carbonates could
850 release comparable quantities of isotopically-heavy CO_2 via calcsilicate formation and
851 decarbonation of dolomite into periclase and calcite, and intrusion into anhydrite could
852 release comparable quantities of S as SO_2 (96). Interaction between petroleum-bearing
853 inclusions and host rock salt could in addition generate halocarbons, estimated at $1 - 3 \times 10^{17}$

854 mol CH₃Cl (35). Intrusion into coal-seams could result in explosive interaction resulting in
855 large-scale coal combustion (93).

856 The Siberian Traps magmas contain anomalously high quantities of S, Cl, F (27),
857 estimated to result in total intrusive and extrusive degassing of 0.2-0.25x10¹⁸ mol S, 0.1-
858 0.25 x10¹⁸ mol Cl, and 0.4-0.7 x10¹⁸ mol F (31). The ultimate source of these volatiles is
859 likely to be assimilation from sedimentary host rocks.

860

861 *6.2.4 Carbon injection perturbations*

862 Given the episodic nature and likely variability in isotopic composition of carbon additions
863 from Siberian Traps volcanism and contact metamorphism, we consider size, rate, and
864 isotopic composition of carbon additions via sensitivity studies within plausible ranges.

865 The effect on $\delta^{13}\text{C}_{\text{carb}}$ of a carbon addition of isotopic composition $\delta^{13}\text{C}_{\text{carb}}$ is given
866 approximately by mass-balance with the surface (atmosphere and ocean) carbon reservoirs
867 (78, 79). The effect on pH is additionally dependent on the rate of addition. We show in
868 Table S10 model results for combinations of addition size, isotopic composition and rate,
869 constrained to result in $\delta^{13}\text{C}_{\text{carb}} = -3 \text{ ‰}$, demonstrating the additional non-linear effect of
870 land-surface weathering feedbacks ('land carbonate compensation').

871 The decline in $\delta^{13}\text{C}_{\text{carb}}$ over the P-T boundary is faster than can be accounted for by
872 a decrease in land carbon burial alone, implying an additional pulse of isotopically light
873 carbon contribution $\sim -2 \text{ ‰}$ to the decline. We show in Figure S7 the constraints from the
874 $\delta^{11}\text{B}$ data on carbon additions over the 50kyr interval 251.95 – 251.90 Ma. The results
875 demonstrate that providing the rate of carbon addition is relatively slow as here, the pH

876 constraints are in fact relatively weak, and are consistent with a range of input isotopic
 877 compositions $\delta^{13}\text{C}_{\text{inj}}$.

878 Table S10 Carbon injection perturbations, constrained to result in $\delta^{13}\text{C}_{\text{carb}} = -3 \text{ ‰}$ when
 879 applied to steady-state condition CO_2Lo

Size (mol C)	Duration	Rate (mol / yr)	$\delta^{13}\text{C}_{\text{inj}}$	ΔpH
4.75×10^{17}	10^5 yr	4.75×10^{12}	-50	-0.07
2.47×10^{18}	10^5 yr	2.47×10^{13}	-10	-0.28
3.32×10^{17}	10^4 yr	3.32×10^{13}	-50	-0.18
1.75×10^{18}	10^4 yr	1.75×10^{14}	-10	-0.64

880
 881 The acidification event at $\sim 251.89\text{Ma}$ requires a rapid addition of carbon on a timescale \leq
 882 10 kyr, with an isotopic composition $\delta^{13}\text{C}_{\text{inj}} \approx 0 \text{ ‰}$ in order to leave $\delta^{13}\text{C}_{\text{carb}}$ unaffected, and
 883 size $\geq 2 \times 10^{18}$ mol in order to produce a sufficient decline in pH and $\delta^{11}\text{B}$ (Figure S8).
 884 Larger (or more rapid) carbon additions produce relatively little additional response in
 885 ocean pH, as the majority of the input carbon remains in the atmosphere, and the response
 886 of the $\delta^{11}\text{B}$ is also non-linear at low pH. The rate of the inferred carbon addition is ~ 3 times
 887 the estimates above for organic carbon from the combined effects of sills ($0.16 - 0.4 \times 10^{18}$
 888 mol C over 10 kyr) and pipes ($0.1 - 0.3 \times 10^{18}$ mol C over ~ 6.5 kyr), but is not unreasonable
 889 given the carbonate carbon source and large uncertainties in these estimates.

890

891 *6.2.5 SO₂ injection perturbation*

892 Contact metamorphism of evaporites resulting in large SO₂ release could potentially also
 893 contribute to the acidification event. Quantitatively the effect (per mol) on ocean pH is

894 approximately twice that of a carbon addition, and given the relative abundance of
895 evaporites to carbonates is therefore likely to represent a smaller contribution to
896 acidification.

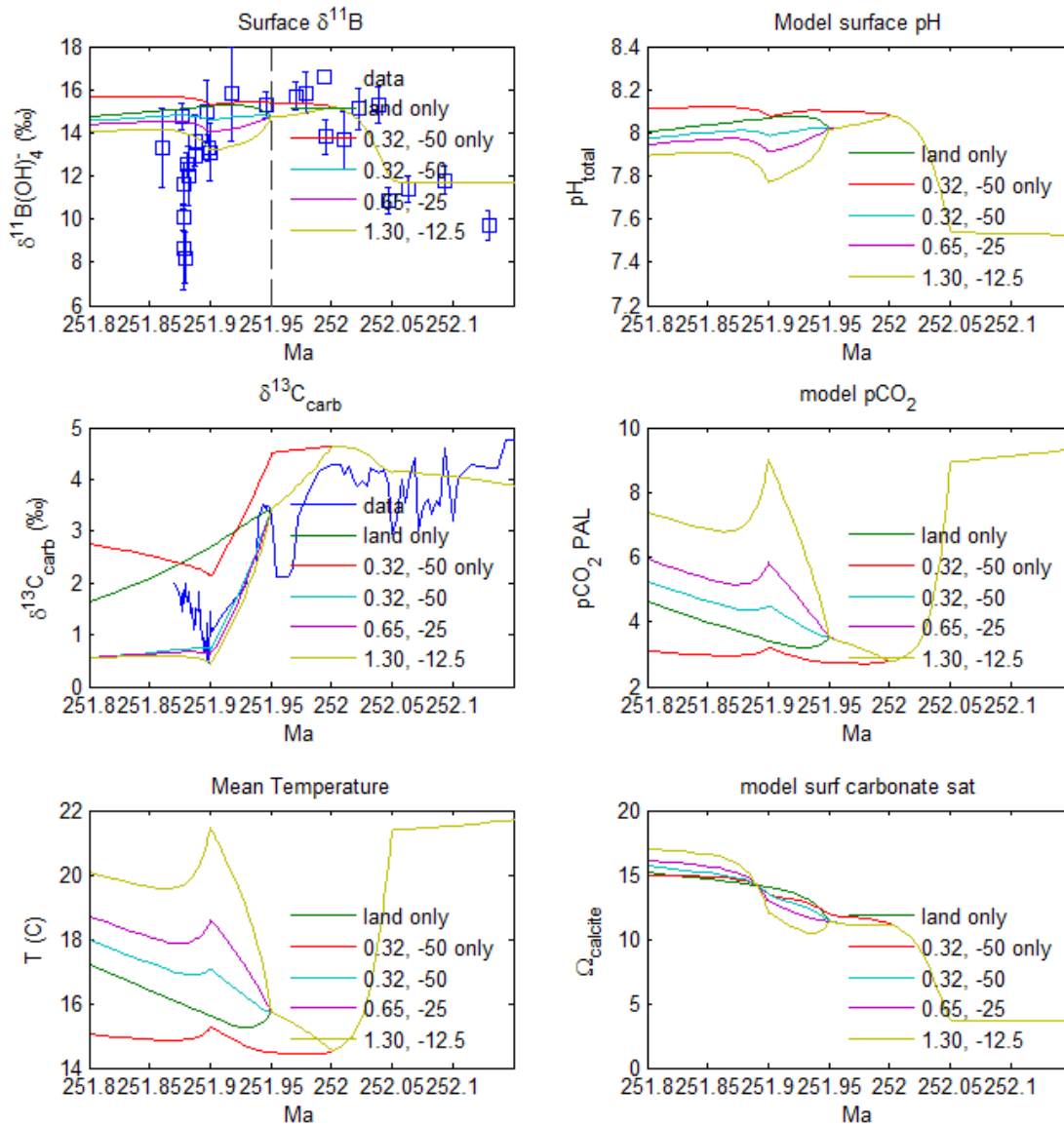


Figure S7 Effect of changing size and isotopic composition of CO₂ input perturbations applied to scenario *CO₂Hi* over 50kyr interval 251.95 – 251.90 Ma. 'land only' shows the effect of interruption in land organic burial, '0.32, -50 only' shows the effect of carbon addition only. Other lines show combined effect of land burial and carbon addition. Perturbations are constructed to result in $\delta^{13}\text{C}_{\text{carb}} \approx -2$ ‰, and correspond to inputs of: 0.32×10^{18} mol / -50 ‰; 0.65×10^{18} mol / -25 ‰; 1.30×10^{18} mol / -12.5 ‰

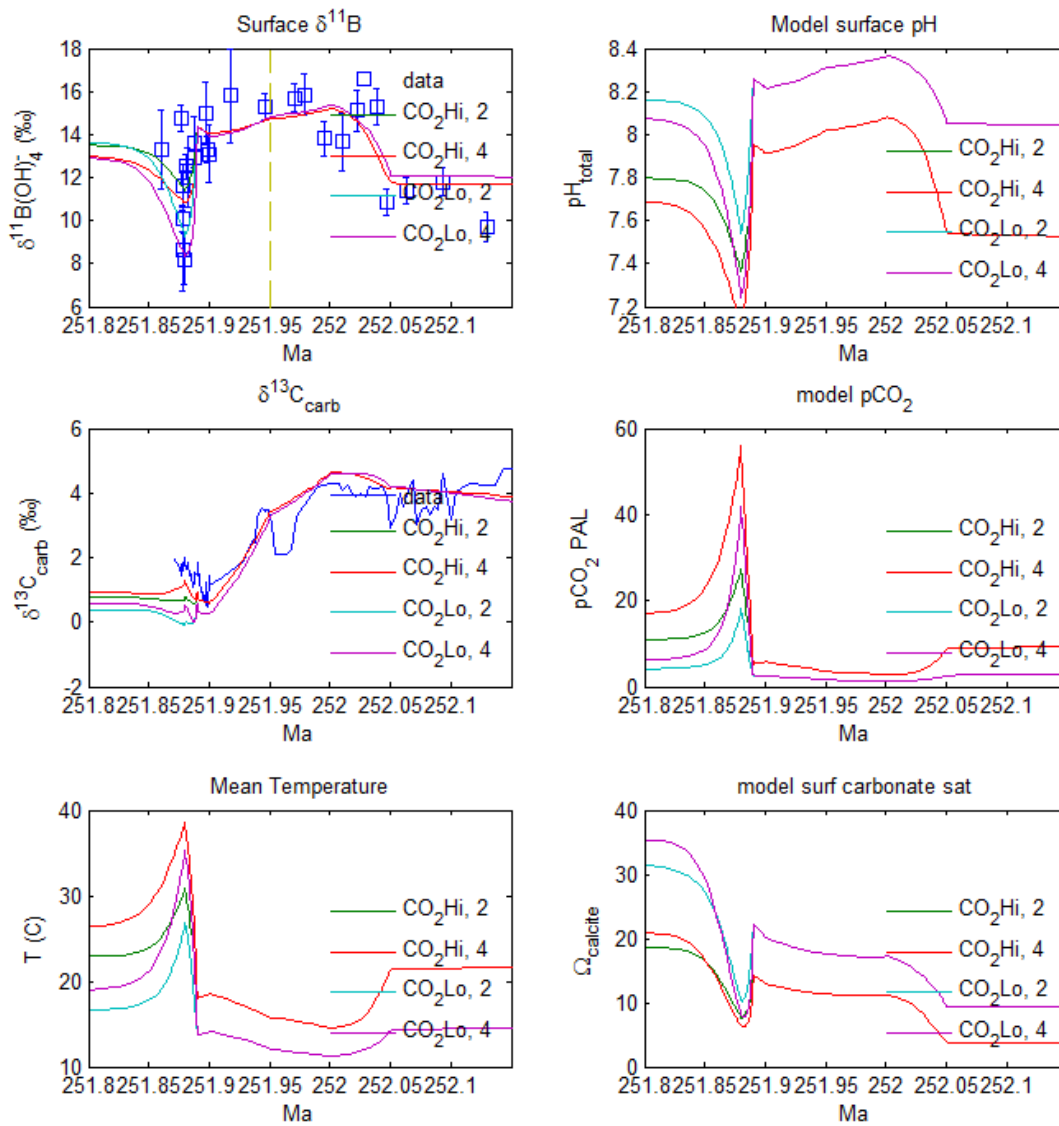


Figure S8 Effect of changing size of CO₂ input perturbations applied to scenarios CO₂Hi and CO₂Lo over 10kyr interval 251.89 – 251.88 Ma. Perturbation sizes are 2x10¹⁸ mol and 4x10¹⁸ mol. Isotopic composition is 2.65 ‰, ie equal to that of sedimentary carbonate.

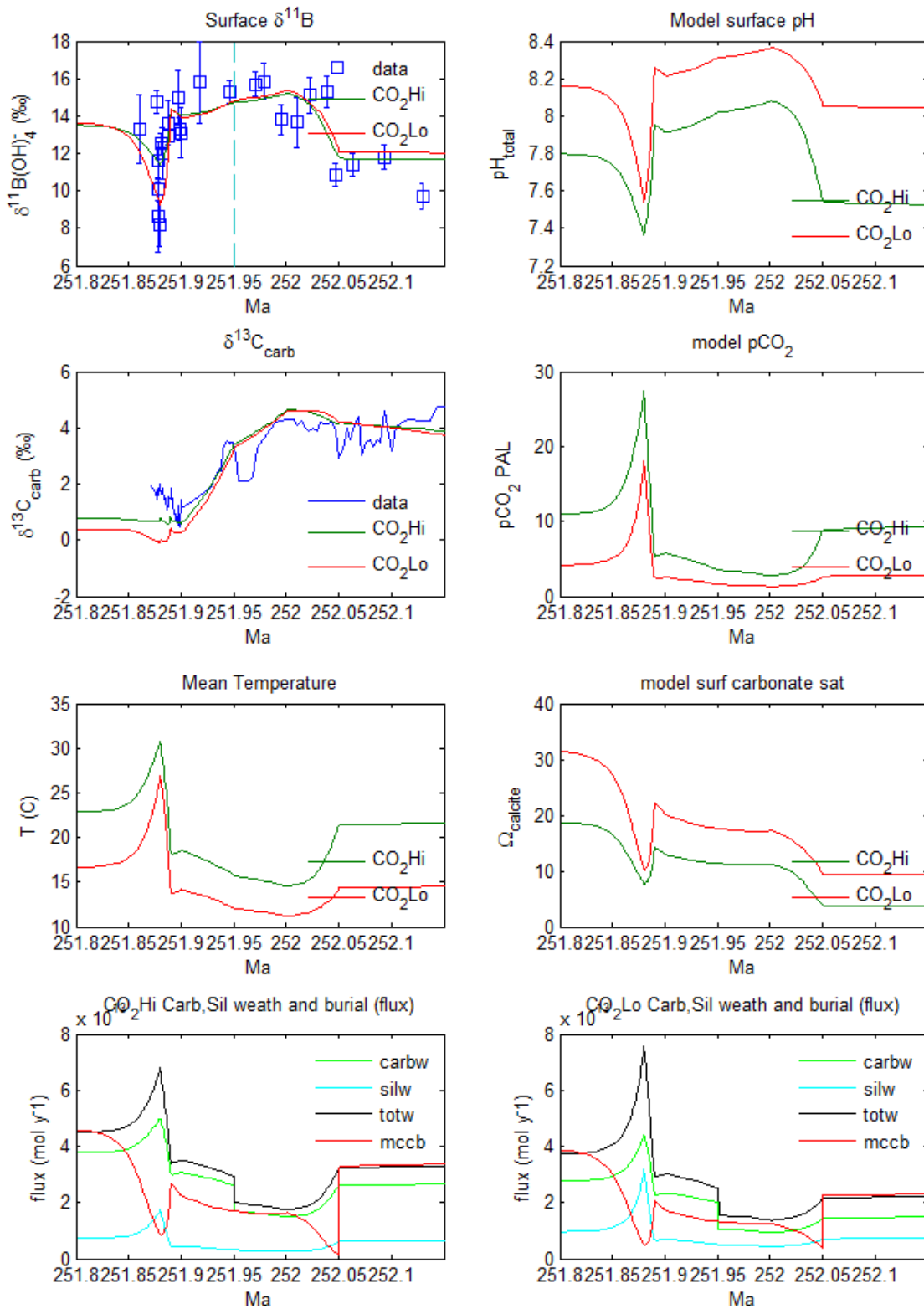


Figure S9 Additional model output for CO₂ scenarios shown in main paper Figure 3. Weathering fluxes are shown for scenario CO₂Lo only.