

when variations in the stable carbon isotope (denoted as $\delta^{13}\text{C}$) in terrestrial plant leaves are taken into account. These factors suggest that the linkage between $p\text{CO}_2$, global warming and the end-Triassic mass extinction remains intact.

The stability of the atmospheric carbon reservoir across the Triassic–Jurassic boundary is inferred from $p\text{CO}_2$ estimates derived using measurements of the $\delta^{13}\text{C}$ composition of pedogenic carbonates, ranging over 20 million years (between the Late Triassic and Early Jurassic), and a palaeosol $p\text{CO}_2$ barometer⁴. A temporal resolution of 20 million years is, however, lower than in the study of fossil leaves³ and so is unlikely to be compatible with detecting the same transient carbon-cycle event. Moreover, a well-characterized feature of the palaeosol $p\text{CO}_2$ barometer is its sensitivity, not only to the concentration of CO_2 in the soil, as acknowledged by Tanner *et al.*², but also to the isotope composition of terrestrial organic matter ($\delta^{13}\text{C}_{\text{OM}}$)^{4,5}. It is therefore necessary to measure $\delta^{13}\text{C}_{\text{OM}}$ coexisting with soil carbonates^{4,5}, but unfortunately, under the very conditions in which palaeosol carbonates form (well-drained soils from arid and semi-arid regions), soil organic matter is rarely preserved, so this feature of the method is often poorly constrained.

Tanner *et al.* calculate $p\text{CO}_2$ by applying a single $\delta^{13}\text{C}_{\text{OM}}$ value because they lacked organic materials with a stratigraphic resolution that corresponds to all of their soil samples. For the interval across the Triassic–Jurassic boundary, however, at least one land-plant stable-carbon-isotope record is available³, providing a means to account for natural variations in $\delta^{13}\text{C}_{\text{OM}}$ at this time. This negative isotopic excursion has subsequently been recorded in organic matter from the Queen Charlotte Islands in British Columbia⁶ and from St Audries Bay, UK⁷, as

well as in marine carbonates and organic matter from Hungary⁸, indicating that it could be a general geochemical fingerprint of the carbon-cycle perturbation across the Triassic–Jurassic boundary.

Calculation of palaeo- $p\text{CO}_2$ levels using the mean isotopic ratios of Triassic and Jurassic palaeosol carbonates determined by Tanner *et al.*, but constrained using stratigraphically detailed terrestrial-plant $\delta^{13}\text{C}_{\text{OM}}$ records from Jameson Land, east Greenland³, indicates a substantial increase across the Triassic–Jurassic boundary (Fig. 1a). Within the margins of uncertainty of the palaeosol approach, this increase in $p\text{CO}_2$ is consistent with the original $p\text{CO}_2$ reconstruction based on fossil stomata of *Ginkgo* cuticles³ (Fig. 1b) and equates to a rise of around 1,000 p.p.m.v. The rise is more strongly expressed if CO_2 -related global warming is allowed to influence the fractionation of soil CO_2 in the reaction-diffusion model in a manner that is consistent with theory (Fig. 1a). Both the stomatal and palaeosol reconstructions (Fig. 1) show a marked $p\text{CO}_2$ increase that is broadly coincident with the emplacement of the Central Atlantic Magmatic Province (CAMP) flood basalts⁹.

Neither the palaeosol nor the stomatal approach to estimating $p\text{CO}_2$ variations in deep time are certain, but both provide self-consistent results for the Palaeozoic⁴ era and the mass-extinction event across the Triassic–Jurassic boundary (Fig. 1). Such self-consistency argues against strongly coordinated ocean–atmosphere buffering of CO_2 from volcanic outgassing during the end of the Triassic, possibly because of the geologically short duration (about 0.5 Myr)¹⁰ and high intensity of CAMP activity. Precise dating of the proposed CO_2 excursion is critical in this context.

Moreover, it is hard to envisage a mechanism other than extreme, CO_2 -forced

global warming that can explain the widespread selective floral extinctions and turnovers that are evident in the fossil record at this time. The proposed transgressive–regressive change in sea level² would be unlikely to provoke such a marked response from the terrestrial biota. The mechanisms that drive mass extinction in the marine and terrestrial realms warrant urgent investigation in the face of the current exponential rise in the Earth's atmospheric CO_2 concentration.

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- Hallam, A. & Wignall, P. B. *Mass Extinctions and their Aftermath* (Oxford Univ. Press, Oxford, 1997).
- Tanner, L. H., Hubert, J. F., Coffey, B. P. & McInerney, D. P. *Nature* **411**, 675–677 (2001).
- McElwain, J. C., Beerling, D. J. & Woodward, F. I. *Science* **285**, 1386–1390 (1999).
- Royer, D. L., Berner, R. A. & Beerling, D. J. *Earth-Sci. Rev.* **54**, 349–392 (2001).
- Cerling, T. E. *Glob. Biogeochem. Cycles* **6**, 307–314 (1992).
- Ward, P. D. *et al. Science* **292**, 1148–1151 (2001).
- Hesselbo, S. P. *et al. Geology* (in the press).
- Pálfi, J. *et al. Geology* **29**, 1047–1050 (2001).
- Marzoli, A. *et al. Science* **284**, 616–618 (1999).
- Hames, W. E., Renne, P. R. & Ruppel, C. *Geology* **28**, 859–862 (2000).

Triassic–Jurassic atmospheric CO_2 spike

question the claim by Tanner *et al.*¹ that atmospheric CO_2 levels remained constant across the Triassic–Jurassic boundary on the grounds of problems with stratigraphic completeness and contamination with atmospheric methane. Because methanogenic CH_4 has a light isotope composition and oxidizes readily to CO_2 , methane–clathrate dissociation and oxidation events cannot be detected by palaeobarometers that use the carbon-isotope composition of palaeosol carbonate.

The palaeosol isotopic data of Tanner *et al.* are not presented in the context of measured sections, nor is there any other indication of stratigraphic completeness that is suitable for identification of a transient greenhouse effect. It is therefore unsurprising that the few sampled palaeosols missed the brief CO_2 greenhouse effect inferred from the stomatal index of fossil leaves in three beds within two continuous sequences in Sweden and Greenland². The duration of the isotopic excursion that occurred at the same time as this transient CO_2 greenhouse effect was probably no more than 500,000 years³. This slim target is easy to miss when examined from a distance of 200 million years.

A more serious problem with isotopic palaeobarometers of CO_2 is the compromising effect of massive dissociation events

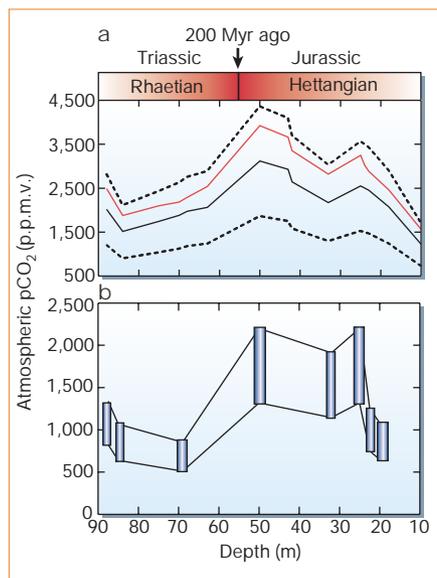


Figure 1 Palaeo-atmospheric $p\text{CO}_2$ changes across the Triassic–Jurassic boundary 208 million years ago. **a**, Atmospheric $p\text{CO}_2$ changes were calculated using the carbon-isotope composition of North American fossil soils² and land-plant organic-matter records from Jameson Land, east Greenland³, using a standard diffusion model⁵, where $S(z)$ (the concentration of CO_2 contributed by biological respiration) is 5,000 p.p.m.v. The horizontal axis represents the depth of successive fossil-plant-bearing beds of the Cape Stewart formation from Jameson Land. $p\text{CO}_2$ values were calculated using a constant Triassic palaeosol carbonate value up to the Triassic–Jurassic boundary and thereafter a constant Jurassic palaeosol value (both from ref. 2). Red line indicates the effects of CO_2 -driven global warming on the temperature dependence of soil CO_2 fractionation^{4,5}, with $S(z) = 5,000$ p.p.m.v. only. Upper and lower broken lines denote the variability in these estimates by varying $S(z)$ between 3,000 and 7,000 p.p.m.v. **b**, Atmospheric $p\text{CO}_2$ changes reconstructed from the stomatal characteristics of fossil leaves from Jameson Land³. Vertical bars denote the upper and lower range for any given depth calculated using this technique.

from methane hydrate reservoirs, because CH₄ oxidizes in the atmosphere within 7–24 years to CO₂, which retains a very depleted carbon-isotope composition (typically –60‰, but as low as –110‰ δ¹³C_{org})⁴. For example, unusually depleted carbon-isotope values for Early Triassic organic matter can be taken as indications of a methane-dissociation event⁵. An Early Triassic CO₂ greenhouse effect is shown by the very low stomatal index of fossil seed ferns⁶, but pedogenic carbonate isotopic palaeobarometers indicate low atmospheric CO₂ levels in the Early Triassic⁷. The pedogenic–isotopic palaeobarometer was not designed for methanogenic isotopic compositions, which compromise this palaeobarometer during several greenhouse transients⁶, and perhaps also at the Triassic–Jurassic boundary.

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1. Tanner, L. H., Hubert, J. F., Coffey, B. P. & McInerney, D. P. *Nature* **411**, 675–677 (2001).
2. McElwain, J. C., Beerling, D. J. & Woodward, F. I. *Science* **285**, 1386–1390 (1999).
3. Ward, P. D. *et al. Science* **292**, 1148–1151 (2001).
4. Whiticar, M. J. in *Atmospheric Methane* (ed. Khalil, M. A. R.) 63–85 (Springer, Berlin, 2000).
5. Krull, E. S. & Retallack, G. J. *Bull. Geol. Soc. Am.* **112**, 1459–1472 (2000).
6. Retallack, G. J. *Nature* **411**, 287–290 (2001).
7. Ekart, D. D., Cerling, T. E., Montañez, I. P. & Tabor, N. J. *Am. J. Sci.* **299**, 805–827 (1999).

Tanner replies — Both Beerling and Retallack question our conclusion that atmospheric CO₂ levels remained relatively stable across the Triassic–Jurassic boundary, partly on the basis of insufficient stratigraphic resolution. However, the stratigraphy of the formations we studied is well known¹. The Lower Jurassic McCoy Brook Formation of the Fundy basin overlies the North Mountain Basalt, which was extruded during the initial stages of volcanism in the Central Atlantic Magmatic Province (CAMP), and so this formation post-dates the main eruptive episode. Palaeosols in this formation occur within 10 m of the formation base; the age of these palaeosols is therefore constrained by the basalt to within several hundred thousand years of the Triassic–Jurassic boundary, which is located several metres below the basalt. Because the duration of the eruptions of CAMP volcanics, which occurred in several pulses, has been established as roughly 600,000 years², palaeosol formation in the McCoy Brook Formation is contemporaneous with the latter stages of the CAMP eruptions. These palaeosols may therefore be expected to record the cumulative effects of the eruptions.

Our palaeo-CO₂ values are calculated from the diffusion-reaction model, which

requires measurements or assumptions for a variety of factors that control soil CO₂, not least the isotopic composition of plant-derived organic matter (δ¹³C_{OM})^{3,4}. Ideally, organic matter in the palaeosol that contains the carbonate is used to obtain this value because the δ¹³C of C₃ plants is known to vary significantly in contemporaneous soils from differing climatic regimes³. Unfortunately, the McCoy Brook Formation, from which the Lower Jurassic carbonate samples were obtained, lacks well-preserved plant material.

The only record of δ¹³C_{OM} across the Triassic–Jurassic boundary available at the time of our calculations provides data for locations in eastern Greenland and southern Sweden⁵, but the negative δ¹³C excursion at the boundary shown in the eastern Greenland data, represented primarily by a single data point, is not apparent in the southern Sweden data set. Moreover, the sediments that contain plant fossils at the eastern Greenland location accumulated under the influence of a significantly more humid climate than existed in the Fundy basin¹, a fact that renders isotope data from this location inapplicable to the interpretation of data derived from the semi-arid palaeosols.

For these reasons, we chose not to use these data, and instead assumed a single value for both the Late Triassic and Early Jurassic that is consistent with published measures of organic matter in Upper Triassic formations⁵ that were deposited under climatic conditions similar to those of the studied formations. This potential source of inaccuracy may ultimately be resolved only by location and analysis of organic material in the McCoy Brook Formation.

The fossil stomatal evidence of Beerling and Retallack seems to indicate a several-fold increase in atmospheric pCO₂ across the Triassic–Jurassic boundary^{5,6}. The use of stomatal indices for calculation of palaeo-CO₂ levels is based on experiments in which modern plants were grown at pCO₂ values of up to twice present levels⁷, but calls for an extrapolation of the experimental pCO₂ values to the much higher palaeo-CO₂ values interpreted for the past. The use of these indices also requires the assumption that the floral response to these conditions was quantitatively the same 200 million years ago as it would be today. This approach may therefore also generate inaccuracies.

As mentioned by Beerling, data from the marine realm that indicate a significant negative carbon-isotope excursion at the Triassic–Jurassic boundary shed new light on the extinction problem⁸, demonstrating an intense but short-lived perturbation of the global carbon cycle of a magnitude that is not consistent with volcanic outgassing. This is supported by a simple mass-balance

calculation, on the basis of the largest estimate of the volume and volatile content of intrusions of the CAMP⁹, showing that outgassing during the eruptions produced 5.6 × 10¹⁶ mol CO₂ in total, an amount that is equivalent to that in the modern atmosphere.

This volume is insufficient to drive the isotopic excursion of marine carbonates⁸ and is unlikely to have affected the atmosphere of the early Mesozoic era, given the high pCO₂ of the Late Triassic¹. Rapid release of seafloor methane hydrate, as suggested by Retallack, has been implicated in other extinction events, and this release may have been triggered by the effects of CAMP volcanism. Such a release, which is consistent with the magnitude of the isotopic excursion, need not have resulted in greatly increased atmospheric pCO₂ as suggested by Retallack, as evidence exists that much of the methane released by this process would be oxidized within the water column, resulting in a brief interval of ocean anoxia¹⁰ and widespread extinction.

Our conclusion stands: the isotope compositions of pedogenic carbonates fail to indicate a substantial increase in atmospheric pCO₂ as a result of the CAMP eruptions. We are all in agreement that although existing methods of estimating palaeo-CO₂ are inexact, their validity is not mutually exclusive, and also that the cause of the end-Triassic extinction event is uncertain, with environmental degradation resulting from the CAMP volcanism probably being involved. Acquisition of new data by both methods from other locations should resolve this uncertainty; better time resolution will constrain the relationships between the abrupt marine extinctions¹¹, the possibly asynchronous floral turnover¹² and the duration of the CAMP eruptions, which may have lasted for several million years¹³.

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1. Tanner, L. H., Hubert, J. F., Coffey, B. P. & McInerney, D. P. *Nature* **411**, 675–677 (2001).
2. Olsen, P. E., Schlichte, R. W. & Fedosh, M. S. *The Continental Jurassic* (ed. Morales, M.) 11–22 (Mus. North. Arizona, Flagstaff, Arizona, 1996).
3. Cerling, T. E. *Spec. Publ. Int. Ass. Sedimentol.* **27**, 43–60 (1999).
4. Ekart, D. D., Cerling, T. E., Montañez, I. P. & Tabor, N. J. *Am. J. Sci.* **299**, 805–827 (1999).
5. McElwain, J. C., Beerling, D. J. & Woodward, F. I. *Science* **285**, 1386–1390 (1999).
6. Retallack, G. J. *Nature* **411**, 287–290 (2001).
7. Beerling, D. *et al. J. Exp. Bot.* **49**, 1603–1607 (1998).
8. Pálffy, J. *et al. Geology* **29**, 1047–1050 (2001).
9. McHone, J. G. *Geol. Soc. Am. Abs. Prog.* **32**, 58–59 (2000).
10. Dickens, G. *Geochem. Geophys. Geosys.* **2**, 2000GC000131 (2001).
11. Ward, P. D. *et al. Science* **292**, 1148–1151 (2001).
12. Pálffy, J. *et al. Geology* **28**, 39–42 (2000).
13. Marzoli, A. *et al. Science* **284**, 616–618 (1999).

