Deep time perspective on rising atmospheric CO2

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ABSTRACT

The accuracy of CO2 hindcasting using fossil Ginkgo stomatal index is ripe for revision for three reasons: exponential rise in atmospheric CO2 over the past decade, discovery of a Kew herbarium specimen of Ginkgo picked in 1754, and increased sophistication of a pedogenic CO2 paleobarometer as an independent parallel record. Past mass extinctions coincide with revised CO2 spikes of 1500 ppm or more. Increases such as the middle Miocene level of 640 ± 71 ppm expected before the year 2100 resulted in biome shifts, with expansion of tropical forests northward, and of grasslands into deserts. Deep time records from paleosols and from stomatal index reveal that CO2 levels less than 180 ppm and more than 1500 ppm are toxic to the biosphere.

1. Introduction

Atmospheric CO2 from fossil fuel burning, forest fires, agricultural ploughing, and oxidation of methane from ruminants and rice paddies has increased dramatically in the past decade (Ciais et al., 2013). Since observations began in 1958 with 316 ppm, atmospheric CO2 levels at Mauna Loa have continually climbed with ± 2 ppm seasonal variation to a peak of 415 ppm in May 2019 (NOAA, 2020). By the year 2100, atmospheric CO2 concentrations could reach 936 ± 140 ppm, based on RCP 8.5 emission scenario of a heterogeneous world with continued population growth (Meinshausen et al., 2011). Such changes are unprecedented in Quaternary ice core measurements showing fluctuation between 180 and 280 ppm (Lüthi et al., 2008), which can show dramatic stepwise increases within closely related fossil leaves (Roth and Dilcher, 1979). This study is mainly based on herbarium specimens now extending back to 1754 (Fig. 1), but is also well constrained because of non-linear decline in stomatal index of Ginkgo over the past two decades newly presented here (Fig. 2).

Stomatal size also increases with increasing CO2 (Franks and Beerling, 2009a, 2009b), and provides another useful CO2 proxy (Franks et al., 2014), but requires many additional measurements of studied specimens not currently available. Stomatal and cell size also increase with genome size, especially for polyploids (Beaulieu et al., 2008), which can show dramatic stepwise increases within closely related fossil leaves (Roth and Dilcher, 1979). Such dramatic differences in cell size were not seen in our data of Ginkgo and other taxa with similar cuticles, in which cell size differences were only noted at very high levels (> 1000 ppm) of CO2.

A significant refinement of the Ginkgo stomatal index proxy is offered here from herbarium specimens of unprecedented temporal range from 1754 to 2015, when observed CO2 had turned the corner to its current dramatic rise (NOAA, 2020). Our new calibration allows a new reconstruction of atmospheric CO2 over the past 300 million years as evidence of past greenhouse crises (Retallack, 2009a), their effects in biodiversity crises (Peters and Foote, 2002), and their role in biogeographic shifts (Retallack et al., 2016).

2. Material and methods

A recently discovered Kew Herbarium specimen of Ginkgo biloba picked in 1754 is well enough preserved to count stomata, despite fungal damage (Fig. 1). All images counted (Conde, 2016) were...
scanning electron micrographs using an environmental instrument, which did not require metal sputter coating. Such images show papillae more clearly than standard macerated preparations, which is important because there is not a 1:1 correspondence between papillae and cells (Fig. 1). Our 1754 leaf anchors the calibration curve for calculating atmospheric CO₂ from *Ginkgo* stomatal index published previously (Retallack, 2009a), and includes new estimates for the intervening years (Fig. 2). This 264-year record of *Ginkgo* stomatal index turned sharply within the recent two decades of CO₂ rise to constrain its future trajectory (NOAA, 2020). That trajectory has also been guided by greenhouse studies of stomatal index of *Ginkgo* grown under different levels of CO₂, but some of those greenhouse studies are suspect because of unnatural malformation of stomata produced by the experiment (Barclay and Wing, 2016). Other greenhouse studies did not attempt such extreme CO₂ levels (Beerling et al., 1998), and had unrealistic Arctic light regimes, but fall close to extrapolation from herbarium observations (Fig. 2). Data on atmospheric CO₂ for all the years of herbarium leaves studied came from direct observations back to 1958 (NOAA, 2020), supplemented by ice core data back to 1754 (Lüthi et al., 2008).

Curve fitting to the relationship between CO₂ and stomatal index has taken many forms (Barclay and Wing, 2016), but an inverse power function based on the principles of Fickian diffusion is most appropriate, with constants and power varied to fit data (Wynn, 2003). Another key feature of our method is counting images with about 600 cells and 60 stomata in both stomatiferous and astomatic areas as a proxy for total leaf conductance. Counting smaller areas of cuticle with as few as 5 stomata gives unacceptable systematic errors of stomatal index: about ± 20% depending on whether 5 or 6 stomata are accidently in the image. This may account for high errors from counting 5–18 stomata by Barclay and Wing (2016). The stomatal index CO₂ palaeobarometer estimates atmospheric CO₂ (C in ppm) from stomatal index (I in % from Eq. (1)) of fossil plants with preserved cell impressions or cuticle in which number of stomata (nₛ) and number of epidermal cells (nₑ) in the same area can be counted. This inverse relationship (Eq. (2)) has an algebraically simplified equivalent (Eq. (3)) between *Ginkgo* stomatal index (I) and atmospheric CO₂ (C). Standard deviations (1σ) of CO₂ concentration (S_C) were calculated by Gaussian error propagation (Eqs. (4) and (5)). Eq. (5) is the partial derivative of Eq. (3) (Wolfram Alpha, 2019).

\[
I = 100 \times \frac{nₛ}{nₛ + nₑ} \quad (1)
\]

\[
C = 239.7 + \frac{1}{2.75255 \times 10^{-7} \times I^{1.79}} \quad (2)
\]
Fig. 3. Estimated atmospheric CO₂ from independent evidence of paleosols and Ginkgo stomatal index, including two high-CO₂ paleosols of the Triassic, Chinle Group in New Mexico (Cotton and Sheldon, 2012), and five low-CO₂ paleosols of the Miocene, Railroad Canyon Formation in Idaho (Retallack, 2009b).

\[
C = 239.7 + 3, 633, 000 \times I^{-4.79}
\]

(3)

\[
S_r = \sqrt{\frac{\Delta C}{\Delta I}} = -17, 402, 070 I^{-0.79}
\]

(4)

\[
\frac{\Delta C}{\Delta I} = -17, 402, 070 I^{-0.79}
\]

(5)

3. Independent validation

CO₂ calculated from our new transfer function (Eq. (3) above) can be checked against estimates of CO₂ from a revised model of carbon isotopic fractionation in paleosols using new paleosol-specific productivity estimates (Breecker and Retallack, 2014), which considerably improves the correlation of paleosol and stomatal index measures (Beerling and Royer, 2011). Our newly calibrated estimates of atmospheric CO₂ from Ginkgo stomatal index are compared with paleosol estimates for the same geological times in Fig. 3: two high-CO₂ paleosols of the Triassic, Chinle Group in New Mexico (Cotton and Sheldon, 2012), and five low-CO₂ paleosols of the Miocene, Railroad Canyon Formation in Idaho (Retallack, 2009b). These are the only two sets of paleosol estimates available using this new method (Breecker and Retallack, 2014). These independent lines of evidence confirm new CO₂ estimates from stomatal index lower than previously thought (Retallack, 2009a), and in line with other CO₂ proxies, such as marine alkenones, marine boron isotopes, and liverwort carbon isotopes (Royer, 2010). Our best estimate for the early Lutetian (50 Ma) of 847 ± 235 ppm CO₂ (supplementary Information Table S1) is lower than an estimate from lacustrine nahcolite+halite of 2055 ± 930 ppm (Lowenstein and Demico, 2006), which may be unduly high because of respiration, either within water of the ancient lake, or within the Gyspoid soil of the dry lake bed (Breeker and Retallack, 2014). Another discrepancy is the Changhsingian (252 Ma) spike with our stomatal index estimate of 2109 ± 1267 ppm for maximum CO₂ of the past 30 Ma, but modelled from marine carbon isotopic data as 2800 to 8300 ppm CO₂ (Cui et al., 2014).

Our new transfer function gives a refined perspective on the history of atmospheric CO₂, which has fluctuated dramatically in deep time (Fig. 4). The new data are comparable with sedimentary mass balance estimates of CO₂ (Berner, 1997), if degraded by a 5 point running average to approximate the 10-million-year steps of mass balance modelling (Retallack, 2001). These new data include more points, and are topologically very similar with the data of Retallack (2009a), but with a y axis of less than half the magnitude of that earlier calibration.

The Ginkgo record goes back only to Late Triassic (226 Ma). Lepidopteris has similar stomatal index to Ginkgo at two localities in South Africa, and related plants with identical cuticles can be used to extend the record back 300 Ma to latest Carboniferous (Retallack, 2009a). Some species of Lepidopteris have been considered vines (McElwain et al., 2007), rather than trees like Ginkgo, but other species of Lepidopteris were dominant plants of low-diversity woodland vegetation (Retallack, 2002). Each ancient species used to infer CO₂ was different from modern Ginkgo in some way, but the uniformity of their stomatal morphology is remarkable (Retallack, 2009a). About 60% of the past estimates of CO₂ in Fig. 4 and Table S1 are relatively low (between 180 and 600 ppm), within the limits of the calibration curve (Fig. 2). As expected, the lowest levels are in the Permian and Pleistocene ice ages (Fig. 4). The spread of deserts and ice caps, marked by widespread Aridisol and Gelisol paleosols at those times (Retallack, 2007), may have reduced global carbon sequestration to a minimum, but could not go lower because of continued CO₂ emissions from volcanoes and global respiration.

4. Greenhouse spikes

Numerous transient spikes in CO₂ (Fig. 4) are now well controlled stratigraphically not only by improved dating of Ginkgo leaves, but by evidence of warm-wet spikes in sequences of paleosols, and in carbon isotopic time series (Retallack, 2009a). Our result may be contrasted with another pilot study proposing values “below 1000 ppm for most of the Phanerozoic, from the Devonian onwards” (Franks et al., 2014), but that result for the Late Triassic, for example, was obtained by averaging estimates from 10 different fossils and localities, including one of 1585 ppm CO₂. The rapidity of both onset and decline of paleoclimatic spikes is confirmed by carbon isotopic composition of organic matter within varved shales, which show greenhouse onset and decline within millennia (Retallack and Jahren, 2008). These Late Permian shale varves are known to be annual because they contain abscised deciduous leaves of Glossopteris. Other confirmation of rapid onset of greenhouse spikes comes from the paleoprecipitation proxy of depth to Bk horizon in 3718 paleosols in Utah and adjacent states over the past 300 million years with sufficient resolution to show very short term (< million
year) transients (Retallack, 2009a). The outlier deep calcic paleosols are not experimental, nor observational errors, because deep-calcic paleosols can be traced as stratigraphic markers throughout the Colorado Plateau of North America (Retallack, 2009a).

Many extreme CO2 transients coincide in time with eruption of large igneous provinces, such as the Late Permian Siberian Traps (Ivanov et al., 2013) and Late Cretaceous Deccan Traps (Renne et al., 2015). Such unusually large eruptions may have created transient greenhouse CO2 spikes and carbon isotopic excursions by thermogenic cracking of coal and carbonate shales (Retallack and Jahren, 2008), or release of methane clathrates (Krull et al., 2000). Radiometric dating of the Late Permian Siberian Traps (Ivanov et al., 2013) and latest Cretaceous Deccan Traps (Renne et al., 2015) shows that eruptions were focused in time to less than two million years. Some spikes, such as the Late Cretaceous CO2 transient, coincide in time with asteroid impact, which would have vaporized biomass to CO2 at ground zero, and added to oxidized carbon by promoting forest fires (Retallack, 1996).

Not only do the spikes rise within thousands of years due to atmospheric injection of CO2 or CH4, but they also fall within thousands of years (Fig. 4). Geologically rapid decline coincides in time with spread of more productive ecosystems into arid lands, where they weathered soil of nutrients (including calcium) more deeply (Retallack, 2009a), and also migration of tropical forests and their deeply weathered soils into high latitudes (Retallack, 2010; Retallack et al., 2016). Greenhouse spikes raised temperatures and humidity to promote hydrolytic weathering, plant growth, and thus carbon consumption and sequestration. A previously published quantification of these relationships from paleosols in Utah (Retallack, 2009a), now must be revised to predict that doubling of CO2 will result in 1.1 °C increase in MAT, and 110 mm increase in MAP from paleosol carbonate depth, and 221 mm increase in MAP from paleosol chemistry (equations in Fig. 5). Doubling estimates are often called “sensitivity”, and have not otherwise been estimated for mean annual precipitation. Global sensitivity of temperature to CO2 doubling has been estimated in other studies (Royer, 2010; Franks et al., 2014) as 3 °C, which is twice our estimate for deep time records of Utah (Fig. 5). Comparable multiproxy studies from elsewhere will be needed to determine whether this is real a discrepancy, but there are now reasons to suspect Utah climate is not globally representative. With current planetary warming, mid-latitude regions have been spared sterilizing heat waves of low latitudes (Frölicher and Laufkötter, 2018; McWhorter et al., 2018) and dramatic deglaciation of polar regions (Overland et al., 2014; Hanna et al., 2014; Retallack, 2010).

Of concern in our century of rapidly rising CO2 are biotic effects of greenhouse spikes, and particularly their effects on biodiversity (Retallack, 2007). The best available metric for defining mass extinctions in deep time is the fossil record of shelled marine invertebrates (Peters and Foote, 2002). CO2 transients above 1500 ppm by our estimation appear to be a planetary Rubicon for mass extinctions (Fig. 6) of the Capitanian (Middle Permian, 262 Ma), Changhsingian (Late Permian, 253 Ma), Smithian (Early Triassic, 251 Ma), Rhaetian (Late Triassic, 202 Ma), and perhaps also Maastrichtian (Late Cretaceous, 66 Ma). Doubts about the Maastrichtian come from inadequate data from stomatal index of Ginkgo (Retallack, 2009a) and ferns (Beerling et al., 2002). Thus an earlier documentation of the relationship between magnitude of greenhouse spikes and mass extinction (Retallack, 2007), is also revised here (Fig. 6). A variety of kill mechanisms are exacerbated by unusually high levels of atmospheric CO2: marine anoxia (Song et al., 2014), coral bleaching (Veron, 2008), wetland tree suffocation (Retallack and Jahren, 2008), acid rain (Retallack, 1996), and intense storms (Kim et al., 2014).

5. Conclusions

Coming atmospheric levels of 450–950 ppm CO2 by 2100 due to anthropogenic forcing (Meinshausen et al., 2011) were last seen during the Middle Miocene (Breecker and Retallack, 2014), when transient atmospheric injections of CO2, perhaps from Columbia River Group flood basalts eruptions (Retallack et al., 2016), created modest mammalian and floral overturn (Retallack, 2013). These crises were followed by diversity-enriching polar migrations and expanded carbon sinks from geographical expansion of tropical forests, polar wetlands, and mid-continental grasslands (Retallack et al., 2016). Deep time records provide useful expectations for toxic highs and lows of atmospheric CO2. The message from the past 300 million years is that CO2, essential for photosynthesis, can be harmful to life on Earth in very
Fig. 6. Relationship between estimated atmospheric CO2 (ppmv) and marine invertebrate extinction (% after Peters and Foote, 2002). Late Cretaceous CO2 is inadequately documented, and problematic estimates from stomatal index are shown (Retallack, 2009a; Beerling et al., 1998). See Supplementary Information Online for data.

large (> 1500 ppm) and very small (< 180 ppm) doses.

Acknowledgements


Appendix A. Supplementary material

Supplementary data to this article can be found online at https://doi.org/10.1016/j.gloplacha.2020.103177.

References


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