

Global patterns in leaf ^{13}C discrimination and implications for studies of past and future climate

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Fractionation of carbon isotopes by plants during CO_2 uptake and fixation (Δ_{leaf}) varies with environmental conditions, but quantitative patterns of Δ_{leaf} across environmental gradients at the global scale are lacking. This impedes interpretation of variability in ancient terrestrial organic matter, which encodes climatic and ecological signals. To address this problem, we converted 3,310 published leaf $\delta^{13}\text{C}$ values into mean Δ_{leaf} values for 334 woody plant species at 105 locations (yielding 570 species-site combinations) representing a wide range of environmental conditions. Our analyses reveal a strong positive correlation between Δ_{leaf} and mean annual precipitation (MAP; $R^2 = 0.55$), mirroring global trends in gross primary production and indicating stomatal constraints on leaf gas-exchange, mediated by water supply, are the dominant control of Δ_{leaf} at large spatial scales. Independent of MAP, we show a lesser, negative effect of altitude on Δ_{leaf} and minor effects of temperature and latitude. After accounting for these factors, mean Δ_{leaf} of evergreen gymnosperms is lower (by 1–2.7%) than for other woody plant functional types (PFT), likely due to greater leaf-level water-use efficiency. Together, environmental and PFT effects contribute to differences in mean Δ_{leaf} of up to 6% between biomes. Coupling geologic indicators of ancient precipitation and PFT (or biome) with modern Δ_{leaf} patterns has potential to yield more robust reconstructions of atmospheric $\delta^{13}\text{C}$ values, leading to better constraints on past greenhouse-gas perturbations. Accordingly, we estimate a 4.6% decline in the $\delta^{13}\text{C}$ of atmospheric CO_2 at the onset of the Paleocene-Eocene Thermal Maximum, an abrupt global warming event ~ 55.8 Ma.

biogeochemistry | ecophysiology | fractionation | PETM

Human perturbation of the global carbon (C) cycle is potentially far greater in rate and magnitude than variations in the recent past, pushing predictions of future climate beyond the calibration range of models based on modern and near-modern observations. Robust predictions of future impacts of rising CO_2 require not only extrapolation of ecological patterns along modern environmental gradients but also insights gained from changing ecological patterns at times of high CO_2 and hot climate in the geologic past (1 and 2). Global patterns of variation in leaf carbon isotope ($\delta^{13}\text{C}_{\text{leaf}}$) values potentially record climate-driven changes in modern plant physiology and biogeochemistry. An understanding of factors controlling plant fractionation (Δ_{leaf}) at the global scale will improve interpretations of past changes in climate and ecology recorded in ancient terrestrial sedimentary organic carbon (2). Patterns in $\delta^{13}\text{C}_{\text{leaf}}$ of living plants at the global scale, however, are unresolved in spite of abundant published data at smaller spatial scales.

In living plants, $\delta^{13}\text{C}_{\text{leaf}}$ values reflect the balance of photosynthesis and stomatal conductance and their coupled response to the environment (3). Edaphic factors (e.g., water availability, altitude, temperature) and plant attributes (e.g. phylogeny and leaf traits) can influence $\delta^{13}\text{C}_{\text{leaf}}$ values (4–6). The relative importance of these factors at the global scale is not known, nor is it clear how they might drive the variations in ancient $\delta^{13}\text{C}_{\text{leaf}}$ values

recorded in either terrestrial organic carbon ($\delta^{13}\text{C}_{\text{TOC}}$) or $\delta^{13}\text{C}$ values of plant biomarkers.

During extreme climate events in the past, changes in the isotope ratio of atmospheric CO_2 ($\delta^{13}\text{C}_{\text{atm}}$) reflect perturbations in atmospheric C fluxes, therefore accurate estimates of $\delta^{13}\text{C}_{\text{atm}}$ are central to estimating past climate sensitivity to changes in $p\text{CO}_2$ (7). Values of $\delta^{13}\text{C}_{\text{TOC}}$ from sedimentary rocks are widely used to estimate $\delta^{13}\text{C}_{\text{leaf}}$ of ancient plants and infer changes in $\delta^{13}\text{C}_{\text{atm}}$ (8). However, the offset between $\delta^{13}\text{C}_{\text{atm}}$ and $\delta^{13}\text{C}_{\text{leaf}}$ (i.e., Δ_{leaf}) varies with climate and plant characteristics and such variations must be accounted for when estimating atmospheric C isotope excursions (CIE) from ancient plant-derived C. For example, during the Paleocene-Eocene Thermal Maximum (PETM), a period of rapid global warming 55.8 million years ago (9), the CIEs in terrestrial organic carbon (TOC) and atmospheric CO_2 likely differ because of changes in Δ_{leaf} that accompanied plant community shifts, warming ($\sim 5^\circ\text{C}$), decreasing precipitation and increases in atmospheric CO_2 (2).

In this study, we provide predictive relationships for Δ_{leaf} variability of woody C_3 plants at the global scale from analysis of published $\delta^{13}\text{C}_{\text{leaf}}$ values, plant functional types, biome, climate, and geography. We use these relationships to show how environmental and ecologic change during the PETM influenced plant fractionation and therefore the CIE recorded by fossil leaf waxes.

Results and Discussion

Leaf Carbon Isotope Fractionation. We used Δ_{leaf} in our analyses as it controls for variation in $\delta^{13}\text{C}_{\text{atm}}$ (10). Δ_{leaf} in C_3 plants is a function of the fractionation associated with CO_2 diffusion (4.4%) and photosynthetic fractionation by Rubisco (27%). For C_3 plants, the concentration of CO_2 in the substomatal cavities of leaves (c_i) is directly proportional to Δ_{leaf} when the atmospheric CO_2 concentration (c_a) is held constant (10) (but see ref. 11). In turn, c_i depends on the flux of CO_2 into the leaf, which is largely regulated by stomatal conductance (g_s) and the flux of CO_2 removed from the leaf for C fixation by assimilation (A) (10). Values of Δ_{leaf} generally decrease with reductions in water availability, reflecting a down-regulation of g_s and increased water-use efficiency (WUE) (10, 12). Thus, Δ_{leaf} is a time-integrated measure of WUE, although leaf temperature (T), mesophyll conductance and differences in c_a are potential confounding factors (11 and 13). Confounding effects are reduced, however, because mesophyll conductance is correlated with A (and with g_s) (11 and 14) and trees appear to maintain leaf

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hand, accounting for the uneven distribution of PFTs along environmental gradients may reveal additional environmental controls on Δ_{leaf} . When regression models are analyzed separately for each PFT, small latitude and T effects on Δ_{leaf} are evident for deciduous angiosperms (DA) and evergreen gymnosperms (EG), the PFTs that span the greatest range in latitude and T (*SI Appendix*). For DAs and EGs, increases in latitude or decreases in T result in slightly higher Δ_{leaf} than predictions based on MAP and altitude alone, indicating limitations on carbon assimilation due to low T or irradiance.

We used a multiple regression model with MAP and altitude as predictors to capture the dominant environmental influences on Δ_{leaf} , and explored whether PFTs explain any residual variation in Δ_{leaf} . This approach reveals significant differences in Δ_{leaf} among PFTs (*SI Appendix*), with EGs between 1.0% and 1.5% lower than the others. More strict control of confounding environmental influences on Δ_{leaf} is achieved by limiting PFT comparisons to plants at the same geographic site (Fig. 2; *SI Appendix*). With this constraint, DAs and evergreen angiosperms (EA) have higher Δ_{leaf} than EGs, by 2.7% and 2.2% respectively. PFT effects on Δ_{leaf} are not solely related to phylogeny or leaf habit. These differences may be related to genetic or phenotypic differences in A and g_s , mediated by leaf morphology, hydraulic architecture, rooting depth, leaf T , and/or mesophyll conductance (3, 10, 11, 13). These traits also vary among species *within* the same PFT; future studies should examine this variation.

Deciduous and evergreen species have contrasting nutrient use and retention strategies (23) that may underlie their dominance in different ecosystems. Our results support the notion that WUE differences are also an important component of plant strategies related to resource availability and likely help explain PFT dominance patterns. Lower Δ_{leaf} implies greater WUE of EGs relative to other woody plants, conferring a potential advantage where water is limiting and a limitation on A when water availability is high (24).

Biome Patterns. Mean Δ_{leaf} values calculated by biome type (Fig. 3; *SI Appendix*) differed significantly, with greatest fractionation in tropical rain forests (mean Δ_{leaf} = 23.4%) and lowest in xeric woodland/scrublands (mean Δ_{leaf} = 17.3%), which compare well with biome level Δ_{leaf} patterns predicted by Kaplan et al. (5). We

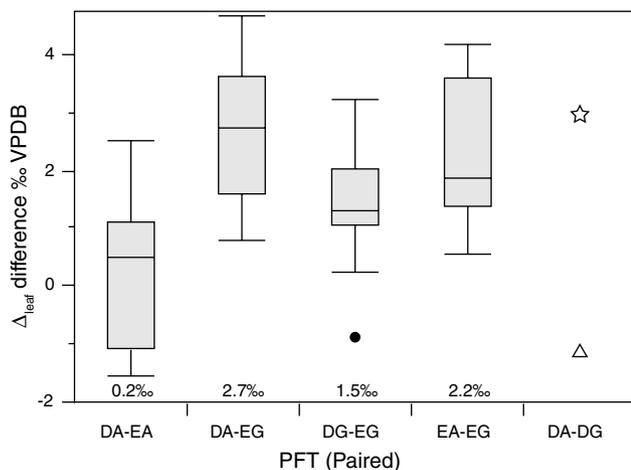


Fig. 2. Differences in Δ_{leaf} between cooccurring plant functional types (PFTs). For each geographic site, mean Δ_{leaf} values of each species were averaged to produce a mean Δ_{leaf} for each PFT. Box and whisker plots show the median, upper and lower quartiles, and maximum and minimum values, with outlier values shown as black dots. Number of sites per PFT comparison are as follows (deciduous angiosperm, DA; deciduous gymnosperm, DG; evergreen angiosperm, EA; evergreen gymnosperm, EG): DA – EA = 16, DA – EG = 17, DG – EG = 21, EA – EG = 12. DA-DG comparison is shown to highlight differences between *Larix* (triangle) and *Taxodium* (star).

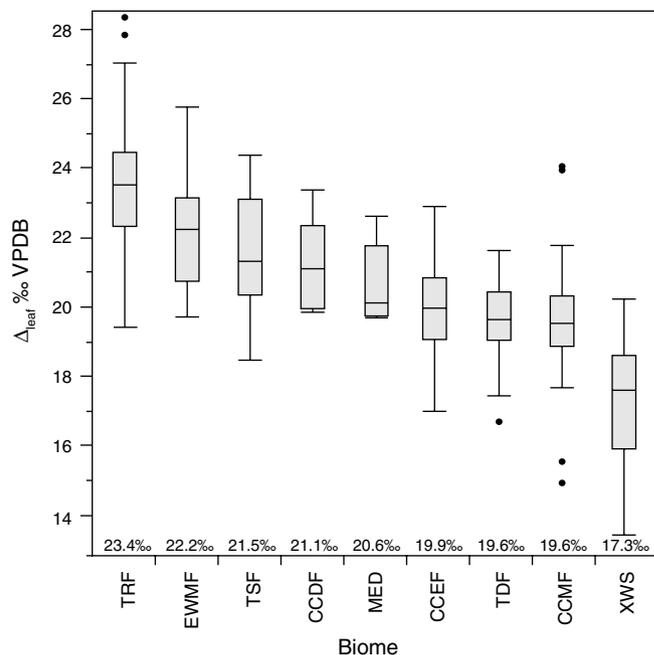


Fig. 3. Box and whisker plots of Δ_{leaf} values by biome. Biome accounts for 66% ($p < 0.0001$) of the variability in Δ_{leaf} in an ANOVA model. Number of samples per biome are as follows (see Fig. 1 caption for abbreviations): TRF = 206; EWMF = 29; TSF = 47; CCDF = 5; MED = 5; CCEF = 53; CCMF = 78; TDF = 26; XWS = 59. Statistical tests of means are shown in *SI Appendix*.

observe a similar pattern when mean Δ_{leaf} for each biome is calculated separately for angiosperms and gymnosperms (*SI Appendix*). Biome type alone explains more Δ_{leaf} variation than MAP (R^2 of 66% and 55%, respectively, *SI Appendix*), because it captures other factors related to Δ_{leaf} , including the spatial distribution of PFTs. PFT influences at the biome scale are readily apparent; cool-cold deciduous forests have higher Δ_{leaf} values (by 1.2%) than cool-cold evergreen forests (Fig. 3; see also Kaplan et al. (5)). The DA-EG differences in Δ_{leaf} (1.0% to 2.7%) calculated for each biome are consistent with results based on residuals of multiple regression (1.0–1.5%; *SI Appendix*) and the paired-site comparison (2.7%; Fig. 2).

Implications for Models. Kaplan et al. (5) recently extended a coupled vegetation-biogeochemical model (BIOME4) to predict Δ_{leaf} at the biome and PFT scale. Our database for C_3 woody plants provides a means for full evaluation of BIOME4's ability to reproduce global patterns in Δ_{leaf} . Since Δ_{leaf} records the time-integrated balance of A and g_s , global patterns in Δ_{leaf} can be used to assess the ability of physiological models to predict the balance between leaf C and H_2O fluxes. Further, incorporating Δ_{leaf} into coupled vegetation-biogeochemistry models have potential to enable predictive mapping of spatial variability in $\delta^{13}C_{\text{leaf}}$ and $\delta^{13}C_{\text{TOC}}$, which could improve studies of the effect of climate on modern and ancient organic matter.

Geologic Implications. The importance of water in regulating Δ_{leaf} values is often invoked in environmental and atmospheric reconstructions from $\delta^{13}C_{\text{TOC}}$ (1). More recently, the role of PFTs in controlling $\delta^{13}C_{\text{TOC}}$ has been identified (2 and 25). Our results show that variation in PFT and MAP independently result in differences in Δ_{leaf} of up to several % (Fig. 1 and 2), and that biome may be a powerful integrator of PFT, rainfall, and other effects on Δ_{leaf} and $\delta^{13}C_{\text{TOC}}$ (Fig. 3). These quantitative relationships can be used to understand temporal and spatial variations in $\delta^{13}C_{\text{TOC}}$ provided that MAP and PFT (or biome) can be estimated in the geologic past.

There are uniformitarian assumptions in using Δ_{leaf} patterns among extant plants to interpret the vegetation of the geological past, however two factors encourage us to proceed. First, the morphological and physiological responses of C_3 plants to water stress are constrained by their fundamental anatomy and biochemistry (26), making it likely that the relationship of water availability and Δ_{leaf} has been similar in direction and magnitude over geological time. MAP and altitude effects on Δ_{leaf} are apparent within as well as among genera, implying that independent lineages respond similarly and quickly. Second, a recent study by Crisp et al. (27) shows that plant lineages rarely shift biomes during evolution, suggesting niche conservatism and supporting the careful use of biome type to constrain Δ_{leaf} values. Nonetheless, applying modern correlations of Δ_{leaf} with environment to the distant geological past may be problematic because of differences in major plant groups prior to the origin of angiosperms in the Cretaceous.

Pollen and leaf fossils document spatial and temporal changes in PFT as well as T and precipitation (9 and 28). Plant biomarkers (chemical fossils), such as n -alkanes (from leaf waxes) or terpenoids (defense compounds) are also preserved geologically (9) and have fewer diagenetic or source effects on their isotopic composition than $\delta^{13}C_{\text{TOC}}$ (29). Plant waxes in ancient sediments reflect productivity-weighted inputs from all PFTs contributing to that deposit. The resulting mixed molecular signal can be used to estimate a $\delta^{13}C_{\text{leaf}}$ value that integrates plant biomass within the catchment. In contrast, terpenoids provide PFT specificity because tricyclic diterpenoids are unique to woody gymnosperms and pentacyclic triterpenoids are unique to woody angiosperms. Therefore, ratios of these terpenoids, after accounting for production and preservational biases, may provide estimates of PFT (30). Also, terpenoid $\delta^{13}C$ values can be used to calculate PFT-specific Δ_{leaf} values (25).

Heterogeneity in rainfall and PFT dominance across ancient landscapes presumably created spatial patterns in Δ_{leaf} just as they do now, even at small spatial scales. To understand differences in $\delta^{13}C_{\text{TOC}}$, $\delta^{13}C_{\text{biomarkers}}$ and the magnitude of CIEs recorded in different regions at the same time, and improve chemostratigraphic correlations using $\delta^{13}C_{\text{TOC}}$, we should account for biome (5), precipitation (9), and PFT (2 and 25) effects on Δ_{leaf} values using relationships like those reported above.

An important goal in studies that document temporal variation in $\delta^{13}C_{\text{atm}}$ is to reconstruct the sources and fluxes of C to the ancient atmosphere (7). $\delta^{13}C_{\text{leaf}}$ and $\delta^{13}C_{\text{TOC}}$ have been used to directly represent $\delta^{13}C_{\text{atm}}$ (4 and 31), but this approach is

fraught with uncertainty (32). It has been suggested that using a simple numerical offset to convert $\delta^{13}C_{\text{TOC}}$ to $\delta^{13}C_{\text{atm}}$ under nonlimiting water conditions is sufficient to control for environmental impacts (4). To the contrary, we argue that secular changes in PFT and climate must also result in significant—but often unaccounted—effects on Δ_{leaf} . Previous studies averaged multiple $\delta^{13}C_{\text{TOC}}$ values to remove variation caused by environmental factors (4, 31, 32), ignoring insights about paleoecology and climate that might be inferred from $\delta^{13}C_{\text{TOC}}$ variations.

We use an example from the PETM to illustrate how our study of modern Δ_{leaf} can improve interpretation of ancient $\delta^{13}C_{\text{biomarkers}}$ and $\delta^{13}C_{\text{atm}}$ (Fig. 4; see *SI Appendix*). Smith et al. (2) documented a shift from a mixed angiosperm-conifer flora (~25% angiosperm) in the late Paleocene to an angiosperm flora (~100%) at the onset of the CIE, and inferred that the 5% decrease in $\delta^{13}C_{\text{leaf}}$ (estimated from n -alkanes) reflected both a decrease in $\delta^{13}C_{\text{atm}}$ (3–4%) and an increase in Δ_{leaf} caused by a shift from conifers to angiosperms (1–2%, (2, 9)). Although the dominant late Paleocene conifers in the Bighorn Basin (*Metasequoia* and *Glyptostrobus* (9 and 33)) were deciduous, they probably had low Δ_{leaf} , as do their extant relatives (34 and 35) (Fig. 2). Yet the PETM in Wyoming is also marked by a change from wetter conditions (MAP ~ 1,380 mm/yr) in the late Paleocene (28) to drier conditions (MAP ~ 800 mm/yr) at the onset of the event (9), which should have decreased Δ_{leaf} . When both PFT and MAP effects on Δ_{leaf} are taken into account (Fig. 4; see *SI Appendix*), the decrease in Δ_{leaf} from declining MAP is countered by the increase in angiosperms, resulting in a net Δ_{leaf} increase of 0.2% and an estimated negative CIE in $\delta^{13}C_{\text{atm}}$ of 4.6%. This estimate of the atmospheric CIE is larger than that calculated by Smith et al. (2), who argued the CIE in $\delta^{13}C_{\text{alkanes}}$ was higher than the atmospheric CIE because of the loss of conifers, and that marine sediments recorded the true CIE. Our Δ_{leaf} correction for both PFT and MAP effects on $\delta^{13}C_{\text{alkanes}}$ suggests the relatively large CIE recorded in the Bighorn Basin and other terrestrial sites (36) may equal the atmospheric CIE. This larger CIE falls within the range recently reported from well-preserved shallow marine carbonate records (–3.5% to –5.0%) and supports the notion that –3.5% is too small (37). A larger atmospheric CIE implies the carbon released at the onset of the PETM was more depleted or greater in mass than presently thought. Adjusting other terrestrial CIE records for biome, MAP, and PFT effects should further refine and reconcile estimates of the magnitude of the CIE.

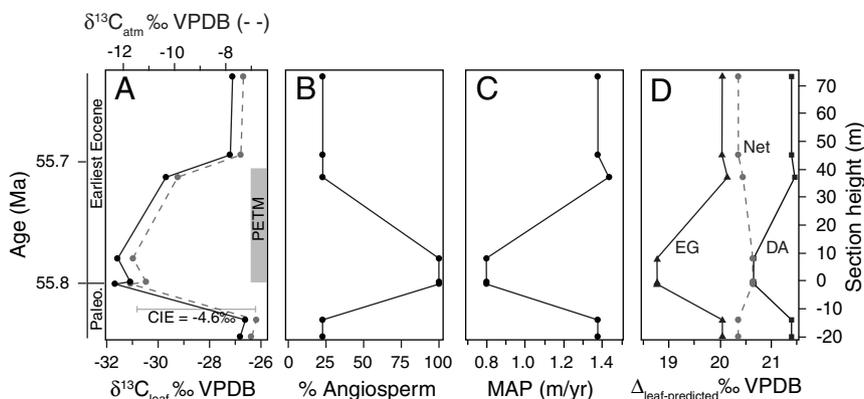


Fig. 4. Conceptual diagram of $\delta^{13}C_{\text{atm}}$ reconstruction for latest Paleocene to earliest Eocene sediments in the Bighorn Basin (WY). The figures are constructed as follows: (A) $\delta^{13}C_{\text{leaf}}$ values as inferred from $\delta^{13}C$ values of n - C_{31} alkanes (2), (B) percent of fossil leaves that are angiosperms, (C) mean annual precipitation (MAP) derived from fossil leaf metrics (9, 28), and (D) modeled net Δ_{leaf} from our Δ_{leaf} -MAP expressions for evergreen gymnosperm (EG; triangles) and deciduous angiosperm (DA; squares) PFTs (see *SI Appendix*). Each model was scaled by % angiosperm to derive a net Δ_{leaf} (dashed line). $\delta^{13}C_{\text{atm}}$ is denoted with a dashed line on A and is derived from the net Δ_{leaf} and $\delta^{13}C_{\text{leaf}}$ values. The negative CIE in $\delta^{13}C_{\text{atm}}$ at the base of the Paleocene-Eocene Thermal Maximum (PETM) is 4.6% (difference between latest Paleocene value and average of initial PETM values). See *SI Appendix* for discussion of calculations and suggested use of these models.

Relationships based on modern plants may not fully quantify the response of Δ_{leaf} to greenhouse climate conditions (high T and $p\text{CO}_2$) such as those of the PETM (12, 13, 38). Given the multiple functions of stomata (water loss prevention, CO_2 provision for A , evaporative cooling), it is unlikely that plants could simultaneously maintain homeostasis with respect to leaf T (15) and c_i or Δ_{leaf} (12) during periods of rapid global change. Studies on Δ_{leaf} of extant relatives of ancient plants under high temperature and $p\text{CO}_2$ conditions and varying degrees of water stress are needed to better interpret changes in $\delta^{13}\text{C}_{\text{TOC}}$ and $\delta^{13}\text{C}_{\text{biomarkers}}$ during the PETM and other events.

Conclusions. We document global patterns in Δ_{leaf} among woody plants, including: (i) a strong positive correlation with MAP ($R^2 = 0.55$), (ii) Δ_{leaf} values 1% to 2.7% lower for EGs than other woody PFTs, and (iii) differences up to 6% in Δ_{leaf} among biomes ($R^2 = 0.66$). The relationship of Δ_{leaf} with MAP shows that the balance of leaf gas-exchange in modern plants is largely driven by water availability, consistent with the role of water in driving global trends in ecosystem-scale primary productivity. By revealing global-scale relationships of MAP, PFT, and biome with Δ_{leaf} , our results uniquely enable applications of these patterns for understanding global-scale processes, such as extrapolation of the distribution of PFTs and leaf carbon and water fluxes under future climate scenarios. In geologic studies, our models of Δ_{leaf} provide tools for interpreting spatial and temporal variation in $\delta^{13}\text{C}_{\text{TOC}}$ and $\delta^{13}\text{C}_{\text{biomarkers}}$, leading to enhanced understanding of paleoecology and atmospheric CO_2 during climate events that serve as analogs of the near future. Our results warn against the assumption in some geologic studies that Δ_{leaf} is invariant in space and time and emphasize that estimating $\delta^{13}\text{C}_{\text{atm}}$ from $\delta^{13}\text{C}_{\text{TOC}}$ or $\delta^{13}\text{C}_{\text{leaf}}$ requires information about biome, PFT, and paleoclimate. We use this approach to produce a refined estimate of the atmospheric CIE (-4.6%) during the PETM and suggest that it be extended to other PETM CIE records to better estimate T sensitivity to greenhouse gases. Such work will aid our

understanding of plant response to extreme climate change both in the geologic past and in the coming century.

Methods

We extracted $\delta^{13}\text{C}_{\text{leaf}}$ values for woody trees and shrubs from 45 publications and one new study (SI Appendix). We excluded $\delta^{13}\text{C}_{\text{leaf}}$ values from juvenile, fertilized, or watered plants, immature or shaded leaves, and understory shrubs. All $\delta^{13}\text{C}_{\text{leaf}}$ values were converted to $\Delta^{13}\text{C}_{\text{leaf}}$ using the Farquhar et al. (10) equation ($\Delta_{\text{leaf}} = (\delta^{13}\text{C}_{\text{atm}} - \delta^{13}\text{C}_{\text{leaf}})/(1 + \delta^{13}\text{C}_{\text{leaf}}/10^3)$) and $\delta^{13}\text{C}_{\text{atm}}$ values estimated for the year of sampling (39) unless reported. Estimating $\delta^{13}\text{C}_{\text{atm}}$ may cause bias in Δ_{leaf} due to seasonal and latitudinal gradients in $\delta^{13}\text{C}_{\text{atm}}$. However, this gradient is $<0.5\%$ during the growing season (40). Species means were calculated for each geographic site to remove within-species variability. For a subset of sites where $\delta^{13}\text{C}_{\text{leaf}}$ values from multiple PFTs were reported ($n = 53$), we calculated paired PFT differences at each site by averaging all species in each PFT and then calculating differences between PFT. For each geographic site, environmental factors were extracted from the publication or derived from global databases (SI Appendix) and biome classifications were assigned based on the BIOME4 model, descriptions in the original publication, and environmental factors. We analyzed the data using linear, least squares regression and ANOVA (SAS JMP 7.0). Δ_{leaf} values and latitude are approximately normally distributed. MAP was \log_{10} -normally distributed and transformed accordingly. Altitude was approximately normally distributed after square root transformation. We performed pairwise comparisons of means using the Tukey-Kramer HSD. Statistics reported within the text and appendices were performed at the site-species combination level, unless otherwise noted ($n = 570$ or less, depending on the model). Similar results were obtained from statistical models at the site level (SI Appendix) using means for all species at a site ($n = 73$ or less).

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