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Geophysical Research Letters

RESEARCH LETTER

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Key Points:

- A multimillion-year delay appears to exist between peaks in low-latitude mountain uplift and the maximum extent of Phanerozoic glaciation
- We describe a mechanism to explain the delay using a combination of climate and geochemical modeling of the periods surrounding uplift
- The mechanism suggests such a scenario potentially happening $10\text{--}50 \times 10^6$ years in the future

Supporting Information:

- Supporting Information S1

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A Capacitor-Discharge Mechanism to Explain the Timing of Orogeny-Related Global Glaciations

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Abstract Over geological timescales, mountain building or orogenesis is associated with increased weathering, the drawdown of atmospheric CO₂, and global cooling. However, a multimillion-year delay appears to exist between peaks in low-latitude mountain uplift and the maximum extent of Phanerozoic glaciation, implying a more complex causal relationship between the two. Here we show that global silicate weathering can be modulated by orogeny in three distinct phases. High, young mountain belts experience preferential precipitation and the highest erosion. As mountains are denuded, precipitation decreases, but runoff temperature rises, sharply increasing chemical weathering potential and CO₂ drawdown. In the final phase, erosion and weathering are throttled by flatter topography. We conclude that orogeny acts as a capacitor in the climate system, granting the potential for intense transient CO₂ drawdown when mountain ranges are denuded; the mechanism suggests such a scenario potentially happening $10\text{--}50 \times 10^6$ years in the future.

Plain Language Summary Over timescales of tens of millions of years or more, plate tectonics can raise large mountain ranges. CO₂ can be removed from the atmosphere through a complex series of processes involving the weathering of rocks, which depends on processes such as rainfall, which in turn is affected by the presence of mountains. The result is that mountain ranges are associated with a reduction of atmospheric CO₂ and global cooling on these very long timescales. An analysis of geological data suggests a multimillion-year delay between peaks in mountain range uplift at low latitudes and the maximum extent of glaciation over the last 400×10^6 years. Our manuscript explains the delay using two numerical or computer models: a climate and circulation model and a geochemical model that simulates weathering processes. We show that weathering, and the implied reduction of atmospheric CO₂, happens most intensely when mountain ranges are eroded, because the ability of weathering to remove CO₂ depends not just on precipitation but also on the temperature of river runoff. Our mechanism intriguingly suggests such a scenario potentially happening $10\text{--}50 \times 10^6$ years in the future associated with projected changes to the height of the Tibetan plateau.

1. Introduction

Over geological timescales, atmospheric CO₂ concentration is stabilized by a negative feedback in which the higher temperatures and precipitation associated with higher CO₂ increase the weathering rate of silicate rocks, ultimately forming marine sedimentary carbonates which sequester atmospheric CO₂ (Maher & Chamberlain, 2014; Raymo & Ruddiman, 1992; Walker et al., 1981). The stable concentration of CO₂ is the result of a balance between the tectonic input rate of CO₂ and its removal by silicate weathering (Bernier, 2004), and it is broadly understood that changes in CO₂ concentration on timescales of 1–100 Myr over the Phanerozoic Eon (the last 541×10^6 years) have been driven by both variation in CO₂ degassing rates and the amplification of silicate weathering due to changes in paleogeography, lithology, and biotic enhancements (Bernier, 2004; Goddérís et al., 2014; Royer, 2014).

The record of Phanerozoic glaciation shows distinct “icehouse” periods in which abundant glacial debris are found (Figure 1, blue bars). For the majority of these icehouse events, the polar “ice line” (e.g., ice cap terminal latitude) has been at $>60^\circ$, while for two major events, the ice line has advanced to $\leq 40^\circ$. The first of these events lasts for approximately 10 Myr and occurs at the apex of the Late Paleozoic Ice Age (LPIA) at approximately 290–300 Myr (Cather et al., 2009; Isbell et al., 2012), while the second is occurring during the present Cenozoic Icehouse. The relatively short duration of these major events is difficult to explain via

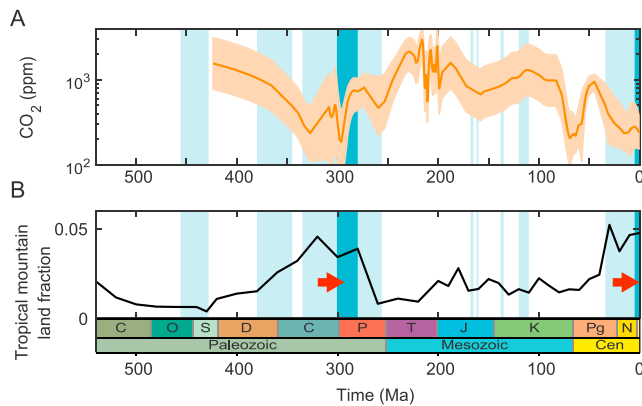


Figure 1. CO₂, glaciation, and low-latitude mountains. Proxy atmospheric CO₂ from combined paleosol measurements (Montañez et al., 2016; Royer, 2014), (a) with uncertainty determined as 16th and 84th percentiles and (b) timeline of fraction of 30°N–30°S land above 1-km elevation (Scotese & Wright, 2018). Both timelines are shown against glacial periods as blue bars (as compiled by Cather et al., 2009), the dark blue representing the most severe glaciation where ice caps reached 40° of latitude. The thick red arrows represent the estimated delay between maximum elevation and maximum glaciation.

changes in the CO₂ input rate, which typically happen over a much longer timeframe (e.g., McKenzie et al., 2016).

Glaciation during the LPIA and late Cenozoic has both been linked to low-latitude mountain building events—the Hercynian and Himalayan orogenies, respectively (Goddéris et al., 2017; Raymo & Ruddiman, 1992)—and in both cases, there is a clear relationship over many tens of million years between mountain building at low latitudes, low CO₂, and cold climate (Figure 1). However, on shorter timescales, it is not clear that either major phases of glaciation or lowest levels of CO₂ correspond directly to the uplift of low-latitude mountains. Rather, there is a suggestion that major phases of glaciations lag behind the maximum extents of high low-latitude mountains by some 10–30 Myr. Figure 1a shows a compilation of atmospheric CO₂ estimates using paleosol isotope measurements from the database of Royer (2014), updated in 2017 to include more recent records (Montañez et al., 2016). It shows that the major ice line advance during the LPIA (Isbell et al., 2012) and the minimum in CO₂ concentration occur at 300–290 Ma, in the Early Permian (Figure 1a), whereas digital elevation models (Scotese & Wright, 2018) and geological assessments (Lardeaux et al., 2001) estimate maximum Hercynian range elevation approximately 20–30 Myr earlier, at around 320 Ma. ⁸⁷Sr/⁸⁶Sr ratios (McArthur et al., 2012) and global sediment abundance (Hay et al., 2006) also suggest a maximum in global erosion at 320 Ma. Additionally, there is a suggestion that an initial minimum in CO₂ concentration at 330 Ma actually precedes maximum mountain extent. It should be noted that while proxy CO₂ measurements are inherently uncertain, the qualitative trend of CO₂ decline from ~320 to 300 Ma is clearly defined beyond this uncertainty (Montañez et al., 2016).

A similar pattern might be inferred over the Cenozoic, although the ongoing nature of glaciation and mountain building, combined with biases that occur when quantifying erosion rates in the immediate geological past (Willenbring & Blanckenburg, 2010) make assessment of the present orogenic cycle difficult. Nevertheless, current peak glaciation and low CO₂ in the Cenozoic do appear to lag significantly behind the peak in low-latitude mountain uplift (Scotese & Wright, 2018; Vérard et al., 2015; See Figure 1). Due to uncertainties in reconstructions and proxies, the precise relationship between the timing of mountain uplift and low-latitude glaciation cannot be proven conclusively. Nevertheless, we suggest that there is sufficient inconsistency between records of low-latitude mountain building, CO₂ concentration, and severe icehouse climates to require a more detailed investigation of how an orogenic cycle affects the key processes that control weathering, especially for what such an investigation may imply for weathering rates and CO₂ concentration in the next few tens of million years, since the long-term future is of great interest to those studying more general aspects of planetary habitability, such as the life span of the biosphere, which has potential implications for the abundance of complex life in the galaxy (e.g., Rushby et al., 2018).

Chemical weathering cannot occur without material supply from physical erosion, as is demonstrated in slowly eroding, transport-limited catchments (West et al., 2005), and approximately half of present-day chemical weathering occurs in and around rapidly eroding mountainous areas (Gaillardet et al., 1999; Hilley & Porder, 2008). Silicate weathering rates are variably dependent on local temperature, runoff rate, and degree of soil shielding (or depth of the weathering zone). Physically based equations for estimating silicate weathering at the catchment scale have been derived, validated against field data (Gabet & Mudd, 2009; West, 2012; West et al., 2005), and applied to present day climate (Hartmann et al., 2014), as well as paleoclimate models having resolutions of approximately 5–7.5° in the horizontal (Goddéris et al., 2017), but have not been coupled to higher-resolution climate models that better simulate orographic enhancement of precipitation, and hence changes in hydrology, over an individual orogenic cycle. In addition, it has been assumed previously that physical erosion is a strong control on chemical weathering rates in mountainous areas (Goddéris et al., 2017), whereas observations and theoretical modeling point to this relationship only being important in lowlands (transport limitation; West et al., 2005; Gabet & Mudd, 2009; West, 2012).

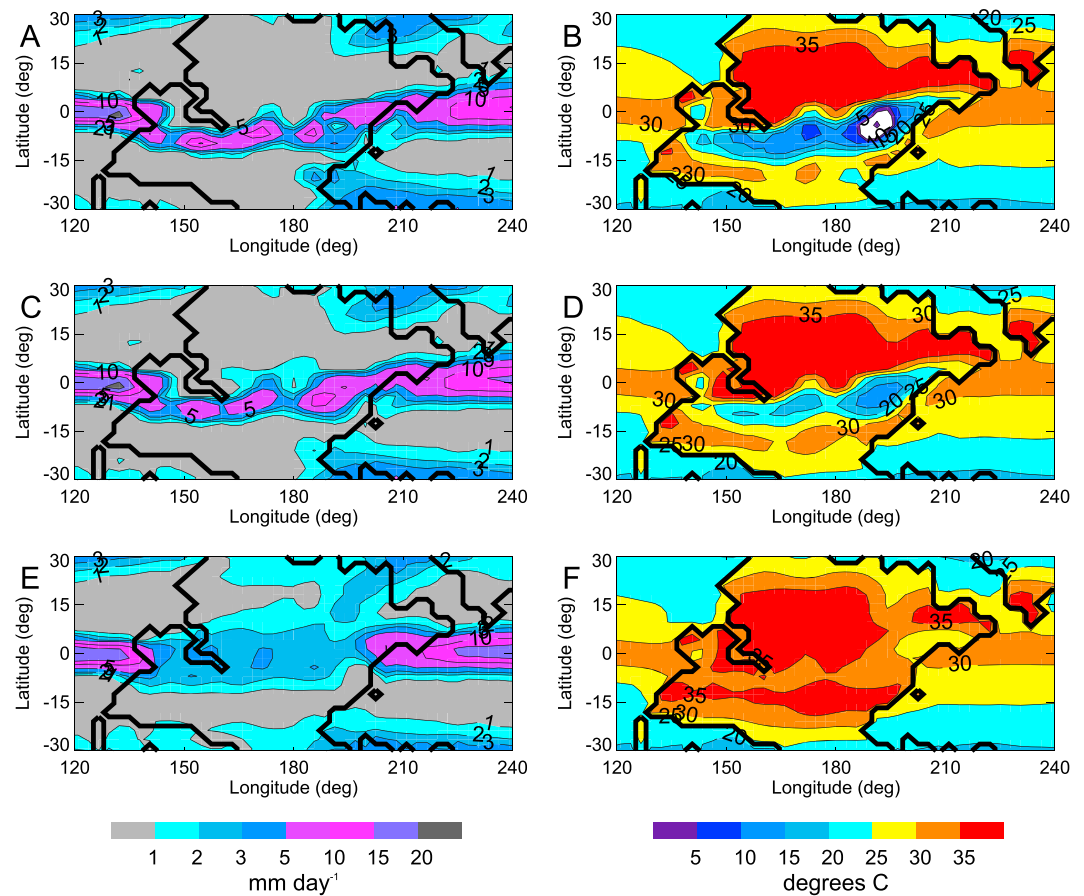


Figure 2. Time-averaged climate in three IGCM4 runs: H (a, b), M (c, d), and S (e, f). (a), (c), and (e) are precipitation (mm/day); (b), (d), and (f) are surface temperature (°C). The land-ocean boundary is shown by the thick black contour.

In this paper we identify a mechanism which links mountain building, weathering, and CO_2 , using the LPJA as a case study. Precipitation, runoff, and erosion are enhanced over orography (Maher & Chamberlain, 2014; West, 2012), but chemical weathering is limited by lower temperatures at high altitudes (Riebe et al., 2004), resulting in three distinct phases for chemical weathering: an increase as mountain building occurs, a decrease when mountain height rises beyond a certain level, and an increase to its highest level tens of million years after peak topographic height, when mountains have been somewhat denuded, providing a significant source of erodible material and higher runoff temperature, while still allowing elevated precipitation. The resulting effect on CO_2 is to produce two dips in concentration—before and after peak topographic height, with the latter dip being the significantly more pronounced one which can allow low-latitude glaciation.

We demonstrate the proposed mechanism using a combination of physical climate and geochemical models. The physical climate is represented by the Intermediate Global Circulation Model 4 (IGCM4) (Joshi et al., 2015), which simulates the large-scale global atmospheric circulation and hydrological cycle, and has been extensively employed in studies of present-day tropical circulation and climate (see Supporting Information S1 for more details). The IGCM4 has a horizontal resolution of 64×128 (“T42”), exceeding the 40×48 (“R15”) resolution of FOAM, which has previously been applied to weathering during the LPJA (Goddéris et al., 2017). The IGCM4 also performs well when compared to present-day observed precipitation, as shown in Supporting Information S1. The model is set up in three configurations (denoted S, M, and H, for short, medium, and high, respectively), each with paleogeography broadly representative of the Permo-Carboniferous (Scotese & Wright, 2018; ~300 Ma; Figure S1), as this is the clearest example of a single low-latitude orogeny. The only difference between each configuration is that the maximum height of the mountain range is varied to represent different states of denudation, with the highest being consistent

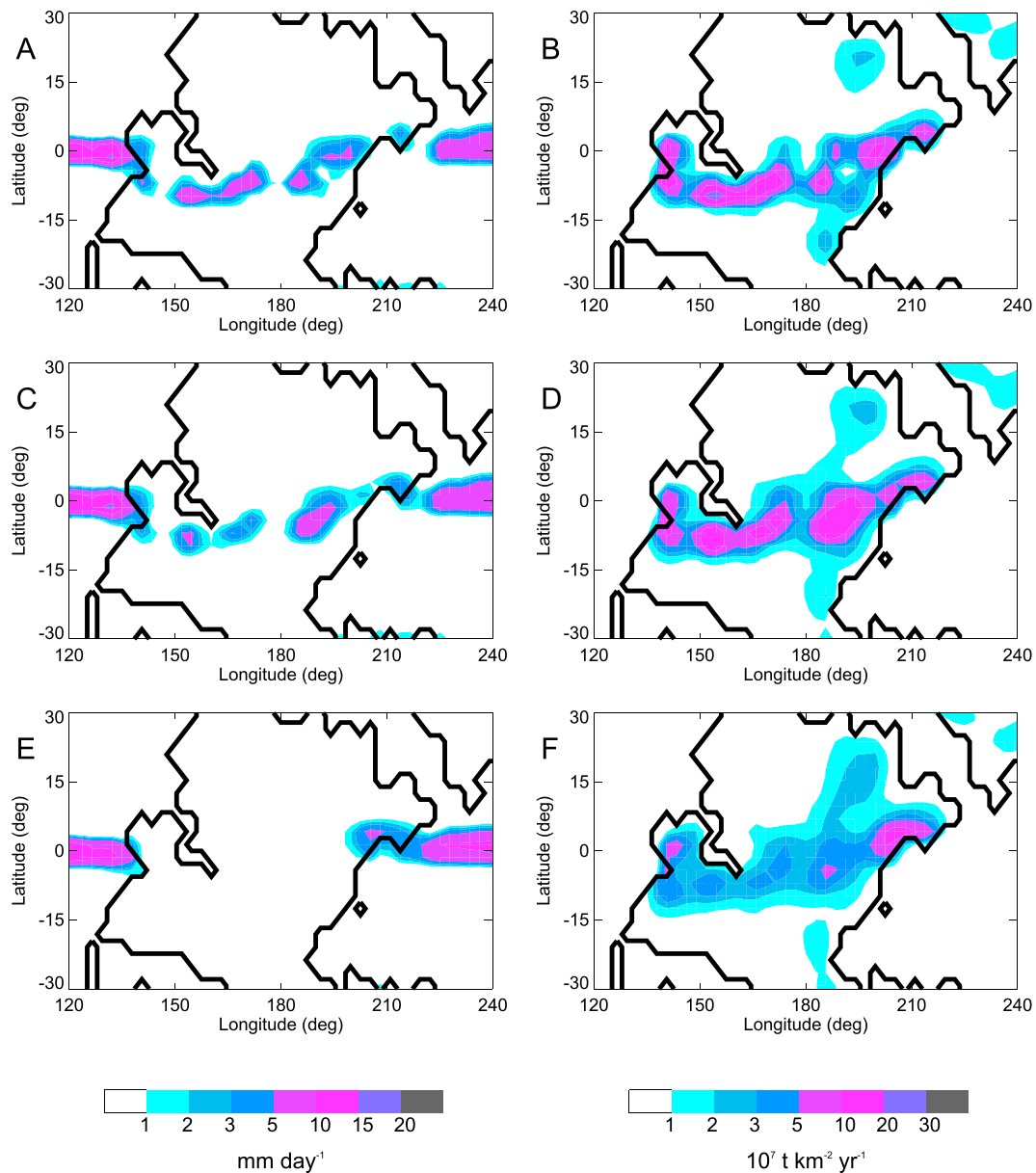


Figure 3. Time-averaged surface hydrology in three IGCM4 runs: H (a, b), M (c, d), and S (e, f). (a), (c), (e) are precipitation minus evaporation (mm/day); (b), (d), and (f) are total silicate weathering ($10^7 \text{ t} \cdot \text{km}^{-2} \cdot \text{year}^{-1}$) over land. The land-ocean boundary is shown by the thick black contour.

with previous work (Godd ris et al., 2017), while the continental configuration remains constant. The present work is therefore isolating the effects of mountain range height on weathering, rather than reconstructing climate history at this time, as was done in Godd ris et al. (2017). The ocean model is a so-called q-flux slab.

Each configuration is run for a variety of CO_2 levels. Silicate weathering rates associated with the changes in climate during an orogenic cycle are parameterized using the model of West (2012) which takes runoff volume and temperature from the IGCM4. The weathering model uses physically based parameterizations of the combined effects of local temperature, runoff, and erosion rates to estimate the chemical silicate weathering rate. We choose this model as, with appropriate parameter choices, it has been shown to fit a broad range of catchment data (Maffre et al., 2018; West, 2012), and we further validate it here against a compilation of present day major world rivers (Gaillardet et al., 1999; see Supporting Information S1). Erosion

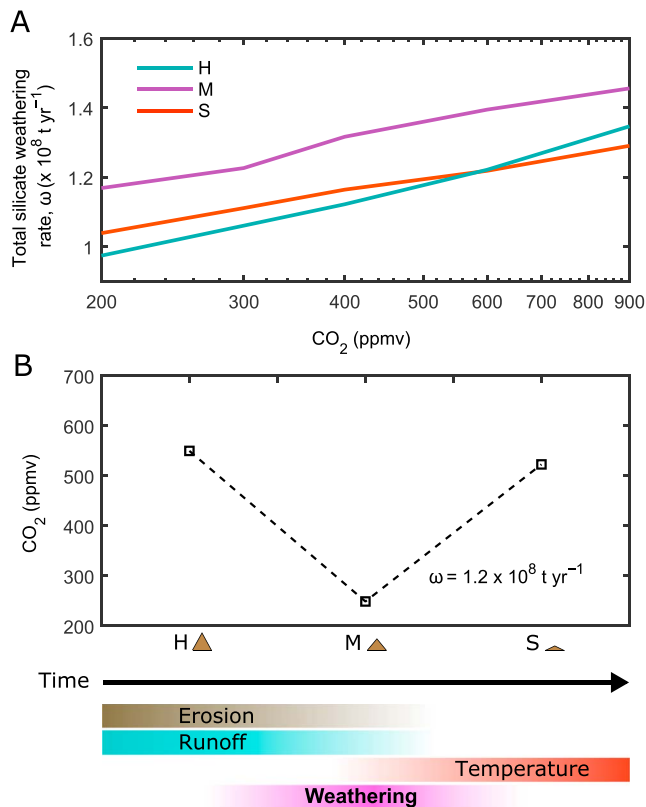


Figure 4. (a) Global chemical weathering rate (in units of 10^8 t/year) versus CO_2 inferred from the weathering model forced by IGCM output (see Methods S1). The mountain configurations are H (blue), M (purple), and S (orange). (b) Timeline of CO_2 concentration required for atmospheric stability assuming a required weathering rate of $1.2 \times 10^8 \text{ t/year}$ in the “denudation” phase, as the mountain range is eroded from H, through M, to S. The colored bars beneath (b) show where runoff (blue), erosion (brown), and runoff temperature or potential chemical weathering (red) are maximized, causing a maximum in total weathering and CO_2 drawdown (purple).

rates are calculated from local elevation using a previously derived global relationship (Montgomery & Brandon, 2002), which relates the erosion rate to mean local relief (see Supporting Information S1). See Supporting Information S1 for more details of both models.

2. Results

Figure 2 shows time-averaged precipitation and surface temperature from the IGCM4 in Runs H, M, and S over the land surface at low latitudes. In all three cases, a band of intense precipitation can be seen at low latitudes, which is analogous to the present-day intertropical convergence zone. While the amount of precipitation over the ocean is similar in all cases, the precipitation over land varies very strongly according to the height of the topography. Runs H and M (Figures 2a and 2b) have intense precipitation over land peaking above 10 mm/day on the flanks of the mountains, while Run S (Figure 2c) has a much weaker and broader band of land precipitation, peaking at 3 mm/day away from the coasts. The most intense precipitation falls on the flanks of the mountains, where the surface temperature is far lower than the sea-level value; in Run H, the most intense precipitation falls onto a surface that is up to 20 K colder than in Run S (Figures 2b and 2f), while in Run M, the intense band of precipitation is in an area that is up to 10 K colder than in Run S (Figures 2d and 2f).

The differences in precipitation and temperature lead to differences in runoff and chemical weathering. Our model run for High mountains (H) has relatively low surface evaporation at the location of the precipitation band because it lies at higher, colder altitudes than in Run M, meaning that the remaining runoff is higher (Figures 3a and 3c). However, Figures 3b and 3d show a clear increase in total silicate weathering flux from Case H to Case M (i.e., mountain denuding) of ~20–50%, despite decreases in both runoff and erosion associated with mountain height. In other words, the effect of higher temperature in Run M dominates the weathering response. As expected, Run S displays even lower rainfall (Figure 2e) and a higher surface temperature, leading to more evaporation, reduced runoff, and reduced chemical and physical weathering

(Figure 3e). The difference in weathering between Runs M and H might in reality be enhanced further by very high erosion rates actually inhibiting weathering (Gabet & Mudd, 2009).

The global implications of these results are examined in Figure 4a, which shows globally integrated weathering rates for the different IGCM runs against atmospheric CO_2 concentration. There is a clear nonmonotonic relationship between weathering and orography: The largest weathering rates for a given CO_2 concentration are in Case M for all values of pCO_2 (purple line). Figure 4b shows the estimated stable level of atmospheric CO_2 for a fixed estimate of the assumed global weathering flux required to balance degassing (taken as $1.2 \times 10^8 \text{ t/year}$) and schematic arrows showing when the key geophysical, climatic, and chemical processes are maximized as the mountain range is eroded away. Equilibrium CO_2 is around ~550 ppmv at maximum mountain height (i.e., Case H) and is reduced to ~250 ppmv after partial denudation (Case M), before increasing back to ~500 ppmv once mountains are denuded completely (Case S). This behavior is consistent with the dynamics of CO_2 concentration between ~320 and ~290 Ma (Montañez et al., 2016; Royer, 2014), during the initial denudation of the Hercynian range.

3. Discussion

Whether peak chemical weathering precedes or lags behind, peak erosion depends on differences in weathering between mountain uplift and denudation periods. Three mechanisms suggest that the former phase permits less integrated chemical weathering than the latter. First, the latter is simply longer-lasting than

the former (Abbott et al., 1997), although it should be noted that uplift timescales are still sufficiently long to allow the silicate weathering feedback to equilibrate atmospheric CO₂. Second, when weathering is kinetically limited (i.e., physical weathering > chemical weathering), there must be accumulation of physically weathered material in the system, either as mobile regolith or as sedimentary deposits (Gabet & Mudd, 2009; Hilley et al., 2010), some of which will be available for subsequent chemical weathering. This contrasts with initial uplift where surface terrains are more likely to need physical erosion before fresh material is available to be weathered. Furthermore, it has been suggested that in such old surfaces, weathering is also inhibited by greater regolith thickness (Goudie & Viles, 2012). When fresh surface is finally exposed, the steep, “spiky” terrain that emerges may strongly limit silicate weathering, due to very high erosion rates, and favor dissolution of carbonate rather than silicate minerals, leading to no net long term CO₂ drawdown (Goudie & Viles, 2012). Third, while glaciers may lead to enhanced weathering through scour-related processes (Anderson et al., 1997), large ice caps will tend to suppress weathering, while retreating glaciers may lead to enhanced carbonate and silicate weathering of glacial deposits (Anderson et al., 1997; Kump & Alley, 1994), suggesting that the net effect of glaciation of high mountains is to pause chemical weathering while accumulating weatherable material.

Figure 1 does strongly suggest that prior to both of the deep glaciations considered here, there is an earlier period of CO₂ drawdown, coincident with initial uplift, but less severe than that following ~20 Myr after peak mountain height, with a rebound of CO₂ in between, associated with the period of peak mountain height. This is consistent with the idea that peak weathering will happen at intermediate erosion rates (Carretier et al., 2014), but the potential for chemical weathering and thus deep glaciation is greater as mountains denude. Present-day global erosion rates in tectonically inactive regions are around 0.1 mm/year (Montgomery & Brandon, 2002), implying 1,000 m of mountain eroded over a timeframe of ~10 Myr. Although the Tibetan Plateau lies at a higher latitude than the LPIA-era Hercynian range, the intense monsoonal precipitation (and weathering) associated with this range is important when considering global weathering (see Figure S6). It is therefore possible that in the next 10–50 Myr, the erosion of the Himalayas, and associated discharge of the weathering capacitor, may overwhelm the negative feedbacks in the global carbon cycle, causing a severe icehouse period to develop in a similar manner to the LPIA.

We conclude that a delayed, climate-controlled capacitor may have allowed low-latitude mountain building to store weatherable silicates at high altitudes which are later discharged, greatly reducing atmospheric CO₂ concentrations and establishing icehouse climates for periods of several tens of million year. The conceptual picture (see Figure 4b) can explain the potential inconsistency between periods of low-latitude mountain building in the geological record and CO₂ drawdown (see Figure 1). It also suggests that CO₂ drawdown may proceed in a transient pulse having a nonmonotonic dependence on mountain height, explaining why such events are difficult to replicate in current zero-dimensional biogeochemical models (Bergman et al., 2004; Berner, 2006). Data sets with enhanced time resolution on these long timescales are needed in order to test this hypothesized mechanism. While our results are most directly applicable to the LPIA, it is possible that future discharge of the weathering capacitor may cause a similar icehouse period to develop in the next 10–50 Myr.

Author Contributions

All authors contributed to devising the study, analysis, and manuscript writing. M. Joshi prepared and performed IGCM simulations, and B. Mills performed weathering rate calculations.

Competing Interests

There are no competing interests.

Data Availability Statement

Relevant model output was placed in an institutional repository at the University of East Anglia (<https://people.uea.ac.uk/en/datasets/search.html>).

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