

How Well Do We Understand the Geological Carbon Cycle?

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Calcite, the largest reservoir of carbon in the crust, in thin section.

1811-5209/25/0022-012\$2.50 DOI: 10.2138/gselements.22.1.12

The foundation of our understanding of the geological carbon cycle, and how this acts as Earth's "thermostat," was articulated in a seminal paper in 1981 (Walker et al). They suggested that silicate weathering on the continents acts as a stabilizing feedback on the carbon cycle such that increased atmospheric $p\text{CO}_2$ leads to increased weathering rates and hence increased removal of CO_2 from the atmosphere. This "textbook model" is at the core of most models of long-term biogeochemical cycles. We summarize evidence that there are many other processes in the geological carbon cycle that may be equally or more important than those in the Walker model. We argue there is a need to move beyond the textbook model in both teaching and research.

Keywords: carbon cycle; feedback; weathering; geological non-uniform Earth system sensitivity

INTRODUCTION AND MOTIVATION FOR THIS ISSUE

"We are so prone to attribute a phenomenon to a single cause, that, when we find an agency present, we are liable to rest satisfied therewith, and fail to recognize that it is but one factor, and perchance a minor factor, in the accomplishment of the total result." Chamberlain (1890)

Carbon dioxide is degassed from the solid Earth into the surface environment (used here to mean the ocean, atmosphere, and biosphere) due to various geodynamic (e.g., magmatism, metamorphism) and surface (e.g., weathering of carbon-rich rocks) processes. As a greenhouse gas, if this CO_2 remained in the surface environment, temperatures would continuously increase and Earth would eventually become uninhabitable, like Venus. Fortunately, processes operate to remove carbon from the surface environment, preventing a runaway greenhouse. The return flux of carbon from the surface environment into the solid Earth comes from a variety of biotic and abiotic processes, such as the deposition of organic carbon-rich sediments and the net removal of carbon into carbonate minerals (Box 1). This cycle of carbon transfer from rocks to the surface environment and back again constitutes the geological carbon cycle (FIG. 1). Operating on 10^5 to 10^6 y timescales, the controls on the geological carbon cycle are generally thought to act as a planetary thermostat, both on Earth and potentially on habitable exoplanets (see *Perspective* by Wordsworth and Minsky 2026 this issue). Because the

residence time of carbon in the surface environment is short (<1 My), differences between the rate of carbon addition to, and removal from, the surface environment over geological timescales could lead to much larger changes in atmospheric $p\text{CO}_2$ than have occurred. Preventing these large changes in $p\text{CO}_2$ and temperature requires a tight policing of the carbon cycle by natural processes that link solid Earth and surface processes, and that are a fundamental part of what makes Earth a habitable planet. Better understanding the tempo and pathways of geological carbon cycling is

of increasing importance as human-induced perturbations dramatically modify Earth's natural carbon cycle and climate. This is especially true as the need for carbon capture and storage mechanisms, many of which mimic natural processes, become more urgent. However, much of how most scientists think about, and teach, this topic is based on a four-decade-old conceptualization that rapidly became the "textbook model." In this issue, we outline the reasons for moving beyond this "textbook model," both in research and education.

THE TEXTBOOK MODEL

In a seminal contribution, Walker et al. (1981) outlined a simple mechanism for climate regulation based on a geological carbon cycle model. The model was built on the concept that *"to maintain a balanced flow of carbon through the surface environment, the average rate of silicate weathering must always have been very nearly proportional to the rate of release of CO_2 by volcanoes and metamorphism"* and that this, in turn, required a negative (i.e., stabilizing) feedback. Because surface processes cannot be the dominant driver of volcanic and metamorphic degassing rates, this feedback must involve the rate of "silicate weathering" being proportional to the amount of CO_2 in the surface environment (Box 1). Walker et al. (1981) suggested that the rate of silicate weathering depends on surface temperature and runoff, which both depend on atmospheric $p\text{CO}_2$ via its role as a greenhouse gas. They also suggested a direct role for $p\text{CO}_2$ in driving silicate weathering rates based on experimental evidence that increases in $p\text{CO}_2$ lead to increased silicate weathering rates. The underlying premise of the Walker et al. (1981) model quickly became accepted by the community, with "silicate weathering" generally assumed to occur solely on the continents; this "continental weathering" feedback is central to most carbon cycle models such that it is widely referred to as Earth's thermostat (e.g., Berner et al. 1983; Sundquist 1991; Zeebe 2012; Lenton et al. 2018; see *Parting Shots* by Kump 2026 this issue).

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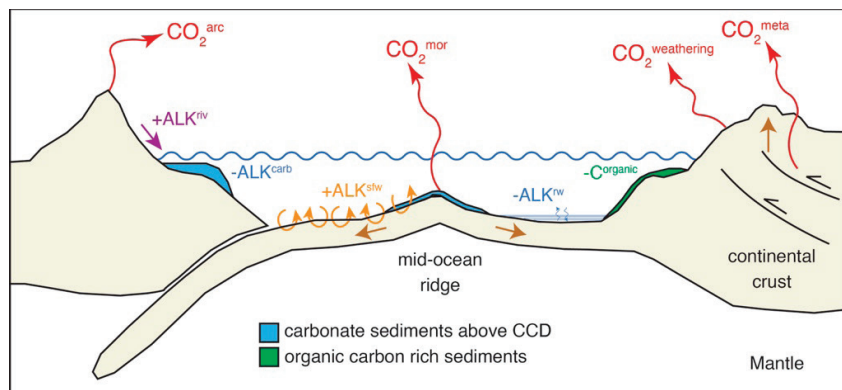


FIGURE 1 Cartoon illustrating the processes critical to the geological carbon cycle discussed in this issue. CO₂ is degassed from the solid Earth (Stewart and Shimizu 2026 this issue) through volcanism, for example, at mid-ocean ridges (CO₂^{mor}) and at arcs (CO₂^{arc}), and is also added to the atmosphere through metamorphic devolatilization (CO₂^{meta}) and weathering of carbon-bearing rocks (CO₂^{weathering}). Carbon is drawn out of the surface environment through burial and preservation of organic carbon in sediments (C^{organic}; Arndt and Hülse 2026 this issue) or formation of carbonate minerals (ALK^{carb}; Rahman and Trower 2026 this issue). The alkalinity needed to form carbonate minerals can come from weathering on the continents (ALK^{iv}; Li and Li 2026 this issue) or seafloor (ALK^{sw}; Bach and Diehl 2026 this issue). However, not all alkalinity added to the ocean acts to remove carbon from the surface environment through carbonate mineral formation, with some lost through reverse weathering (ALK^{rw}; Rahman and Trower 2026 this issue). Brown arrows represent tectonic forcings associated with the uplift of continental rocks and the formation of new oceanic lithosphere.

The textbook carbon cycle model (Fig. 2A) is based directly on the Walker et al. (1981) model, often accompanied by a carbon isotope mass balance to determine the fraction of carbon buried as organic carbon (Box 1). Such thinking is deeply entrenched in much of the scientific community across fields such as Earth history, paleoclimatology, and global biogeochemical cycles, as well as in undergraduate, and even high school, educational programs. For example,

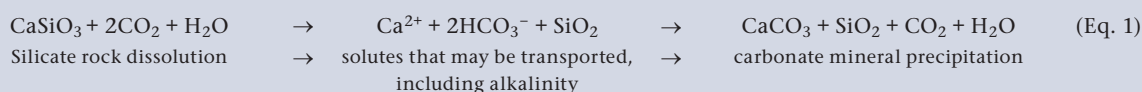
inversion of a variant on the textbook model has been used to predict Phanerozoic pCO₂ levels (e.g., Berner and Kothavala 2001). The model suggests that large changes in pCO₂ were associated with events in Earth history, such as the evolution of land plants and uplift of mountain belts, that changed the strength of the continental silicate weathering feedback. More sophisticated biogeochemical models also link the rate of supply of nutrients (particularly phosphorus) from the continents to the biosphere to the rate of continental silicate weathering (e.g., Jenkyns 2010; Lenton et al. 2018). In these models, biological productivity and organic carbon burial are tied to the rates of solid Earth degassing and continental weathering. The textbook model is also commonly

used as a framework for interpretation of paleo-seawater chemistry. A classic example is the long-standing suggestion that uplift of the Himalaya led to increased silicate weathering and a cooling climate over the last 30–40 My, and that this is recorded in the increase in the Sr isotopic composition of seawater, and other similar proxies, over this time (e.g., Raymo and Ruddiman 1992; Bickle 1996; Li et al. 2021). However, if the textbook model doesn't capture the main controls on the carbon cycle, then models of paleo-pCO₂, nutrient fluxes, and seawater chemistry that are based on it are unlikely to be robust.

Although the “Walker” model of a continental weathering feedback succinctly explained how the Earth could stay habitable for most of its history, its other predictions and implications were not initially thoroughly tested. Instead, this hypothesis has become what Chamberlain (1890) describes as a “controlling idea” into which data are fit and the model molded so as to explain apparently contradictory evidence (or in more modern terminology, confirmation bias). A question readers of this issue may like to ponder is, *would an octopus, when considering why their ocean habitat had remained habitable for billions of years, have ascribed this to processes operating out of sight on the continents?*

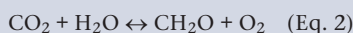
Box 1 REMOVAL OF CARBON FROM THE SURFACE ENVIRONMENT.

Two processes dominate removal of carbon from the surface environment: formation of carbonate minerals and burial of newly formed organic matter. Most carbonate minerals are simply formed by dissolving carbonate rocks on the continents and precipitating carbonate minerals as sediments in the ocean (a process that is generally biologically mediated). In its simplest form, this has no impact on the geological carbon cycle. Carbon degassed from the solid Earth (e.g., from volcanoes and metamorphism; Stewart and Shimizu 2026 this issue) can be taken into carbonate minerals only after transformation of the degassed CO₂ into carbonate alkalinity (HCO₃⁻ and CO₃²⁻). The alkalinity is largely generated by the dissolution of silicate minerals through reactions such as (Urey 1952):



More complicated, but conceptually equivalent, reactions breakdown a range of silicate minerals, producing alkalinity (e.g., Li and Li 2026 this issue). In the textbook model, silicate rock dissolution (weathering) occurs on the continents (e.g., Li and Li 2026 this issue), but this can also occur on the seafloor (ocean crust and overlying sediments; Bach and Diehl 2026 this issue). Additionally, the transportable solutes in Eq. 1 can back-react to form silicate minerals and release CO₂ (reversing the weathering; Mackenzie and Garrels 1966; Rahman and Trower 2026 this issue) such that not all alkalinity generated by silicate weathering is consumed by carbonate precipitation.

Burial of carbon in organic matter can be represented through the imbalance between the forward (photosynthesis) and reverse (aerobic respiration) reactions (Arndt and Hülse 2026 this issue):



Only a small fraction of the CO₂ consumed in the forward reaction is ultimately buried as organic carbon. This burial rate is generally not suggested to be directly related to the CO₂ degassing rate and, thus, does not act as a direct negative feedback. Based on the carbon isotopic composition of carbonate rocks typically being ~0‰, and the carbon degassed into the surface environment being ~-5‰, mass balance suggests that roughly 10%–20% of carbon is buried as organic carbon (~-30‰), with the other 80% buried as carbonate minerals (e.g., Des Marais et al. 1992).

BEYOND THE TEXTBOOK MODEL

Despite its rapid acceptance, Walker questioned his own textbook model just 11 years after it was published (François and Walker 1992; François et al. 1993), but it has remained the cornerstone of most geological carbon cycle models and textbooks (although with notable exceptions, e.g., Edmond and Huh 2003). In the more than 40 years since this model was first described, numerous lines of evidence have come to light that suggest it provides a far from complete view of the geological carbon cycle.

First, studies of continental weathering generally have not found a strong climatic control on weathering rates (e.g., Li and Li 2026 this issue) and even the relative importance of climatic parameters such as temperature, runoff, and vegetation in controlling continental weathering rates remains poorly constrained. For example, Brantley et al. (2023) recently reviewed a wide range of constraints and concluded that the temperature dependence of the rate of continental weathering is only about one-third as strong as that used in most long-term carbon cycle models (e.g., Berner and Kothavala 2001; Lenton et al. 2018).

Second, a wide range of observations suggest that other processes provide direct feedbacks on the geological carbon cycle. These include evidence of: (i) a strong temperature dependence of alkalinity generation via seafloor basalt dissolution (Bach and Diehl 2026 this issue); (ii) early formation of authigenic clays in marine sediments partially “reversing” the weathering that sourced these elements and acting as a sink of alkalinity (Rahman and Trower 2026 this issue); and (iii) a temperature dependence of the release of carbon during weathering of organic carbon-bearing sediments (Soulet et al. 2021). Additionally, many environmental parameters affect the efficiency of organic carbon burial (e.g., Hilton et al. 2015; Arndt and Hülse 2026 this issue). Seafloor basalt weathering works in unison with continental silicate weathering as a negative feedback on the carbon cycle and increases the average

temperature sensitivity of weathering across Earth’s surface (Krissansen-Totton and Catling 2017). In contrast, reverse weathering works against silicate weathering, and organic carbon weathering acts as a source of carbon to the surface environment and a positive feedback on the carbon cycle; i.e., increased temperatures lead to greater CO₂ production from organic carbon weathering if this does not become supply limited (Li and Li 2026 this issue), further increasing surface temperature (FIG. 2B). Many of these additional fluxes are thought to be as large today—or at times in the geological past—as the alkalinity flux generated by continental silicate weathering (e.g., Krissansen-Totton and Catling 2017; Zondervan et al. 2023). Clearly, continental silicate weathering is not the only process involved in regulating the geological carbon cycle.

These differences between the textbook model (FIG. 2A) and a more realistic (but still simplified) geological carbon cycle model (FIG. 2B) show how multiple interacting processes may interact to regulate the geological carbon cycle. The number of processes shown in FIGURE 2B is to some extent arbitrary. Such a figure could be drawn with many more processes, although in our view this figure probably encompasses the most important ones on ~0.1 to 10 million-year timescales. That said, we emphasize that there are large uncertainties about the relative roles of all the processes shown in FIGURE 2B, and that they likely played different roles at different times in Earth history, leading to different responses of the Earth system to CO₂ degassing. The important points about the differences between FIGURE 2A and 2B are not the details, but in the recognition that: (i) there are multiple feedbacks that can be both positive and negative, and that are controlled by a wide range of different factors; and (ii) the ocean plays a fundamental role in the system. Despite the presence of positive feedbacks that may have varied in importance throughout Earth history, these feedbacks always sum to produce a net negative feedback on geological timescales, as predicted by the “textbook” model.

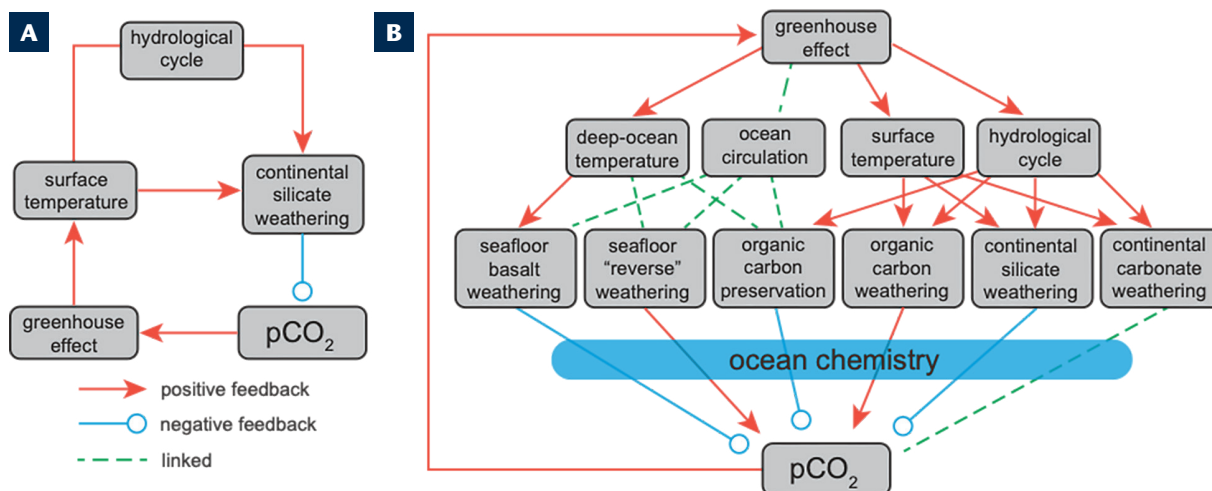


FIGURE 2 Systems diagrams for the geological carbon cycle thermostat operating in the surface environment showing (A) the textbook model (from Kump et al. 1999) and (B) a more complete view of the relevant processes. Positive feedbacks are reinforcing (e.g., increased pCO₂ increases the greenhouse effect, increasing surface temperature), while negative feedbacks are stabilizing (e.g., increased surface temperatures increase seafloor and continental silicate weathering rates, increasing removal of CO₂ from the surface environment and hence drive cooling). Linked processes are relationships where a change in a parameter modulates one or more process that in turn impact CO₂, but without a simple feedback (e.g., pCO₂-driven changes in climate affect ocean circulation, which impacts many aspects of carbon cycling but whether the net result is a positive or negative feedback cannot be readily determined). The more nuanced view shown in (B), while still simplified, recognizes that that carbon

cycle is controlled by multiple interacting processes, which can act as both positive and negative feedbacks on atmospheric pCO₂. It recognizes submarine processes (LEFT) may be as, or more, important than subaerial ones (RIGHT), and that ocean chemistry acts to moderate many aspects of the carbon cycle. For example, because most surface carbon is stored in the ocean, with <2% in the atmosphere at pre-industrial levels, pCO₂ doesn’t simply depend on the mass of carbon in the surface reservoir but how carbon is partitioned between the atmosphere and oceans, which is a function of circulation, temperature, and seawater carbonate chemistry. Both conceptualizations include a net negative feedback that stabilizes the carbon cycle, but in (B), this is not simply continental silicate weathering, and the efficacy of the feedback depends on many aspects of the Earth system, such that the dominant feedback may have varied over Earth history.

Multiple Links and Feedbacks Regulate the Geological Carbon Cycle

The continents are not made of wollastonite. Continental silicate weathering is often simplified to the terms of Eq. 1, where increased continental weathering acts to dissolve Ca- and Mg-rich silicate minerals and the alkalinity released is consumed by precipitating carbonate minerals removing carbon from the surface environment. In reality, many weathering reactions produce residual clays as by-products, and subsequent erosion and deposition of these clays influence the types of reverse weathering reactions and organic carbon burial efficiencies on the marine shelf (Arndt and Hülse 2026 this issue; Rahman and Trower 2026 this issue). Further, many lithologies exist in the continents and increased continental weathering rates can cause fluxes of carbon and alkalinity to change independent of reaction 1. For example, weathering of organic carbon-rich rocks releases CO₂ to the environment (Soulet et al. 2021), weathering of carbonate rocks can lead to CO₂ release or drawdown if the Ca/Mg ratio of the weathered and precipitated carbonate minerals differ (Garrels and Berner 1983), and weathering of sulfide-rich continental rocks generates acidity, which can lead to carbon release from carbonate rocks (Spence and Telmer 2005).

The ocean basins are not lined with Teflon. In the textbook model, continental silicate weathering delivers alkalinity to the ocean where it is consumed through precipitation of carbonate minerals, which removes carbon from the surface environment. However, numerous lines of evidence suggest that the ocean basins are not unreactive vessels that play a passive role in the carbon cycle. Instead, the seawater–mineral reactions at and below the seafloor, an area twice as large as the area of the continents, play important roles in the carbon cycle. For example, the mafic rocks that line the ocean basins, and the overlying sediments, are reactive, and reactions between these and the ocean can both generate alkalinity (Bach and Diehl 2026 this issue) and consume it (Rahman and Trower 2026 this issue).

Linked processes complicate attribution of controlling factors. Many processes associated with continental silicate weathering are linked to other processes that affect the carbon cycle. For example, increased pCO₂ is associated with an invigorated hydrologic cycle coupled to increased silicate weathering rates in the textbook model (FIG. 1A), but these would also be linked to increased nutrient fluxes to the ocean and increased deposition of clay-rich sediments. In turn, there would be increased preservation of organic carbon in sediments (e.g., Hemmingway et al. 2019; Arndt and Hülse 2026 this issue). Indeed, it has been argued that over the Neogene, the primary role of weathering in the Himalaya in the carbon cycle is burial of organic carbon in rapidly accumulating sediments and not alkalinity generation by silicate mineral dissolution (France-Lanord and Derry 1997).

The Moderating Role of Ocean Chemistry

In the textbook model, the ocean is simply a location in which alkalinity derived from continental silicate weathering is lost via carbonate mineral formation removing carbon from the surface environment (FIG. 2A; Box 1). However, it is now clear that many processes affecting both inorganic and organic carbon cycling depend on ocean chemistry and temperature. First, the major ion chemistry of the oceans controls what fraction of the carbon in the surface environment is in the atmosphere, and hence impacts climate. This effect comes in part through the role of seawater major ion chemistry in controlling the activity coefficients relevant to the carbonate system (e.g., Hain et al. 2015; Zeebe and Tyrrell 2019). More directly, the calcium

concentration in seawater controls the carbonate ion concentration at calcite saturation such that, other things being equal (e.g., total dissolved carbon concentration and calcite saturation state), higher seawater calcium concentrations will be associated with higher pCO₂. Thus, while the textbook model directly relates changes in atmospheric pCO₂ to changes in the carbon content of the surface reservoir, this is not necessarily the case. Indeed, it is plausible that the drop in pCO₂ over the Cenozoic may be largely related to changes in seawater major ion chemistry rather than changes in the total carbon content of the surface environment (e.g., Zeebe and Tyrrell 2019; Derry 2022). Further, changes in solid Earth CO₂ degassing rates will lead to changes in atmospheric pCO₂ that are dependent on seawater major ion composition at the time (Hain et al. 2015). Second, the major ion composition of the oceans plays an important role in controlling the thermodynamics of seawater–ocean crust and seawater–sediment reactions that occur at the seafloor (Bach and Diehl 2026 this issue; Rahman and Trower 2026 this issue). Finally, the concentration of key species in seawater, such as dissolved O₂, can control the fraction of organic carbon that is delivered to the seafloor that becomes preserved in the sediment (e.g., Arndt and Hülse 2026 this issue).

A NET NEGATIVE FEEDBACK IS STILL REQUIRED

The fact that Earth has maintained a habitable climate for >3.5 Gy strongly suggests that changes in solid Earth CO₂ degassing are balanced by changes in CO₂ drawdown; i.e., the sum of the positive and negative feedbacks in the geological carbon cycle (FIG. 2B) is always negative on long timescales. To fully understand how this constrains the processes involved in the carbon cycle will require that we better understand solid Earth degassing rates and their variability. Stewart and Shimizu (2026 this issue) show there is almost an order of magnitude uncertainty in the modern rate of solid Earth CO₂ degassing; this directly corresponds to the uncertainty on the CO₂ drawdown fluxes required to balance the geological carbon cycle. Uncertainty in paleo-degassing rates is even higher. Future work aimed at better constraining degassing rates, both in the modern and the past, will be needed to advance existing models of Earth's carbon cycle thermostat.

The co-existence of multiple feedbacks, both positive and negative, as well as linked processes that do not act as direct feedbacks but do lead to changes in carbon cycling, is becoming clearer. Each feedback may operate on different timescales and with differential sensitivities to climate. Further, the relative contribution of a given feedback may depend upon the boundary conditions. A given magnitude change in the CO₂ degassing rate, but that occurred under different climatic, tectonic, biospheric or geomorphic conditions, or with different ocean chemistry, may result in a different net feedback response (e.g., FIG. 3). We refer to this variability in the sensitivity of the Earth system to carbon cycle perturbations at different times in Earth history as the “geological non-uniform Earth system sensitivity” (GUESS). Under some boundary conditions, this sensitivity will be such that the sum of all positive feedbacks to a perturbation may be larger than the sum of all negative feedbacks on short timescales, leading to maximum pCO₂ levels greater than would be expected based on the textbook model. In this case, the timescale of recovery from the perturbation would also be longer than that without any positive feedbacks, but the net negative feedback would still result in a return to the initial conditions given enough time. Under different boundary conditions, the geological Earth system sensitivity could be such that the sum of all feedbacks generates a stronger net

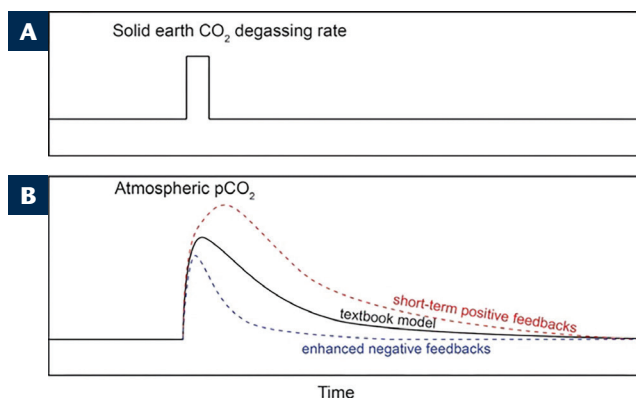


FIGURE 3 Examples of how multiple feedbacks, both positive and negative, can lead to different responses to a perturbation to the carbon cycle. A short-lived increase in degassing rate followed by a return to pre-perturbation conditions (A) would lead an increase in atmospheric CO₂ followed by a gradual return to pre-perturbation conditions in the textbook model (solid line in B). More realistically, the same pulse in degassing may lead to short-lived positive feedbacks (red dashed line) under certain boundary conditions, whereas under others, the negative feedbacks may be much stronger than in the textbook model (blue dashed line).

negative feedback than predicted by the textbook model, leading to lower peak pCO₂ levels and a shorter perturbation (Fig. 3).

TESTING CARBON CYCLE MODELS USING EARTH HISTORY

To assess the relative importance of the processes involved in Earth's carbon cycle thermostat (Fig. 2B), we need models that incorporate all of these processes, as well as robust datasets to test them against (see *Toolkit* by Kirtland Turner and Vervoort 2026 this issue). The transition from a greenhouse to icehouse climate over the Cenozoic, which seems to track a decrease in atmospheric pCO₂ (Fig. 4A), has provided a common test case for such studies. Different processes act to modify fluxes of carbon and alkalinity in different ways and can leave characteristic signatures in the elemental and isotopic compositions of seawater that is then recorded in the marine sedimentary record (Fig. 4). For example, weathering of average continental silicate

rocks produces a flux of alkalinity to the ocean associated with high dissolved ⁸⁷Sr/⁸⁶Sr and ¹⁸⁷Os/¹⁸⁸Os. In contrast, weathering of seafloor basalts or continental basalts produces a flux of alkalinity to the ocean associated with low dissolved ⁸⁷Sr/⁸⁶Sr and ¹⁸⁷Os/¹⁸⁸Os. In both cases, the alkalinity flux leads to removal of carbon from the surface environment into carbonate minerals. In contrast, organic carbon burial is associated with a flux of carbon out of the surface environment, but is not associated with large Sr or Os fluxes. However, organic carbon weathering is associated with both a flux of C into the surface environment and a high ¹⁸⁷Os/¹⁸⁸Os flux due to the strong association of Re with organic matter. Along with other element (e.g., Mg, Ca) and isotope (e.g., ⁸⁷Li, ⁸¹³C, ⁸²⁶Mg, ⁸⁴⁴Ca) proxies, the hope is that the relative roles of different processes in regulating the carbon cycle can be constrained from paleorecords.

Robustly determining the relative importance of different processes involved in Earth's carbon cycle thermostat is challenging for two main reasons. First, this requires constraining carbon and alkalinity fluxes and reservoir sizes today and in the past, both of which have substantial uncertainties (Stewart and Shimizu 2026 this issue; Rahman and Trower 2026 this issue). For example, the only direct constraint on paleo-alkalinity fluxes comes from estimates of the mass of carbonate minerals formed at any time in Earth history. The difficulty in reconstructing such mass estimates is compounded by the fact that it is only the fraction of new carbonate minerals that derive their alkalinity from silicate weathering, and not from carbonate weathering, that directly leads to draw-down of atmospheric CO₂. Second, some of the elemental and isotope fluxes associated with processes and proxies important to the carbon cycle (Fig. 2B) are not well constrained, and the changes in their abundance in seawater are subject to non-trivial uncertainties. For example, the increase in seawater ⁸⁷Sr/⁸⁶Sr and ¹⁸⁷Os/¹⁸⁸Os over the Cenozoic (Fig. 4) could be a record of increased weathering of continental rocks, decreased weathering of seafloor basalts, or changes in the isotopic composition of the rocks being weathered. Likewise, the increase in Mg, and decrease in Ca, in seawater over this time (Fig. 4) may reflect: (i) decreasing ocean crust production rates that lead to a smaller Mg sink and Ca source associated with alteration of ocean crust, (ii) increased silicate weathering on the continents due

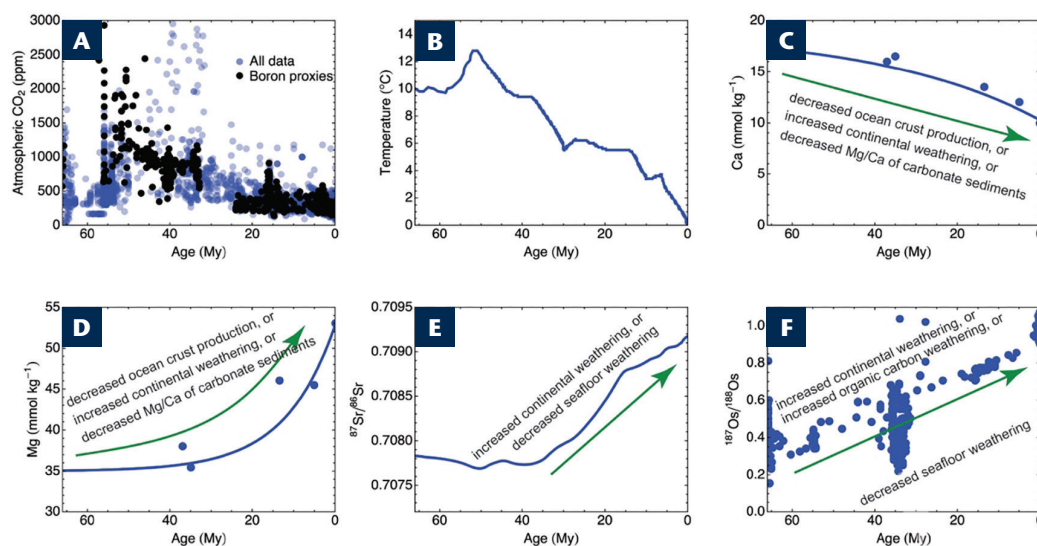


FIGURE 4 Illustration of the kinds of proxy data that can potentially be used to test models of Earth's carbon cycle thermostat. Decreased atmospheric CO₂ concentrations over the Cenozoic (A) are accompanied by decreased deep ocean temperatures (B). Various elemental and isotopic proxies for changes in tectonic processes and weathering fluxes, based on changes in

seawater chemistry, can be explained in multiple ways (C–F, see green arrows and annotations for examples). The blue curves in (C) and (D) are fit through a larger dataset including measurements going back 100 My (Zeebe and Tyrrell 2019). All data (and associated references) are compiled in Coogan and Caves Rugenstein (2025).

to uplift leading to increased fluxes of alkalinity into the ocean and hence removal of Ca, and/or (iii) changes in the fraction of dolomite in carbonate sediments being weathered or deposited. Future models need to not only include a wider range of carbon and alkalinity cycling processes than most extant models, and time-variability in all fluxes, but also need to fully incorporate the uncertainties in the models and data. The articles in this issue provide a starting point for such advances.

LOOKING FORWARD

The seminal contribution of Walker et al. (1981) put forward perhaps the simplest plausible model for a planetary thermostat, that of a single feedback based on the rate of continental silicate weathering, and marine carbonate precipitation, being dependent on atmospheric $p\text{CO}_2$ (Fig. 2A). After decades of research, the evidence for this operating as the sole planetary thermostat is not compelling and the temperature sensitivity of this feedback appears to be weaker than that required to explain paleoclimate reconstructions using carbon cycle models. This has led to the conclusion, reached a long time ago by François

and Walker (1992), that other feedbacks must be important (Fig. 2B) and the net feedback is likely to come from the sum of many processes, operating with different strength at different times in Earth history dependent on the boundary conditions. The articles in this issue highlight paths forward to move beyond teaching the textbook model, and framing interpretations of Earth history this way. Unraveling the full range of mechanisms involved in regulating the geologic carbon cycle should continue to motivate future research. In particular, we hope that future undergraduates will learn a more nuanced view of the long-term carbon cycle (e.g., Fig. 2B). This will better position the next cohort of geoscientists to understand the ways in which anthropogenic perturbations are disturbing the natural system and the mechanisms that humans might borrow from nature to try to undo the harm we are doing.

ACKNOWLEDGMENTS

We thank Sasha Turchyn and Cin-Ty Lee for reviews of this manuscript and Principal Editor Sumit Chakraborty for editorial advice.

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