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**Biogeodynamics: Bridging the gap between surface and deep Earth processes**

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## **Biogeodynamics: Bridging the gap between surface and deep Earth processes**

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### **Abstract**

Life is sustained by a critical and not insubstantial set of elements, nearly all of which are contained within large rock reservoirs and cycled between Earth's surface and the mantle via subduction zone plate tectonics. Over geologic timescales, plate tectonics play a critical role in recycling subducted bioactive elements lost to the mantle back to the ocean-biosphere system, via outgassing and volcanism. Biology additionally relies on tectonic processes to supply rock-bound "nutrients" to marine and terrestrial ecosystems via uplift and erosion. Thus the development of modern-style plate tectonics and the generation of stable continents were key events in the evolution of the biosphere on Earth, and similar tectonic processes could be crucial for the development of habitability on exoplanets. Despite this vital "biogeodynamic" connection, directly testing hypotheses about feedbacks between the deep Earth and the biosphere remains challenging. Here I discuss potential avenues to bridge the biosphere-geosphere gap, focusing specifically on the global cycling and bioavailability of major nutrients (nitrogen and phosphorus) over geologic timescales.

### **Introduction**

Primary productivity is a major driver of elemental cycling at Earth's surface, forming an important link between the geosphere and the biosphere. Autotrophic fixation of volcanically-sourced atmospheric CO<sub>2</sub> into organic matter, coupled to the production of molecular oxygen in its most efficient form, provides important feedbacks to climate, weathering, and the redox state of Earth's surface reservoirs. While CO<sub>2</sub> outgassing and silicate weathering remain the primary controls on atmospheric CO<sub>2</sub> levels over geologic timescales, primary productivity can effectively regulate atmospheric CO<sub>2</sub> in the short-term (e.g., [Falkowski, 1997](#)). In

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3 addition, long-term burial of organic carbon was a prerequisite to the buildup of oxygen in  
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5 Earth's early environments (Berner, 1991).  
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8 Carbon dioxide alone is insufficient to fuel life. Organisms require greater than 20 additional  
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10 major and trace elements to produce biomass and metabolic machinery (e.g., Frausto da Silva  
11 & Williams, 2001). Given the ubiquity of CO<sub>2</sub> at Earth's surface, it is thus the availability of  
12  
13 these "nutrients" that dictates the success of primary producers in natural environments.  
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16 Nitrogen (N) and phosphorus (P), in particular, are ubiquitous in life as we know it and are  
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18 required for the formation of many essential biomolecules, including nucleic acids (DNA and  
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20 RNA) and proteins. These nutrients form the basis of the canonical "Redfield ratio", which  
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22 postulates that marine organic matter is produced and recycled at a nearly constant C:N:P  
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24 ratio of 106:16:1 (Redfield et al., 1963), following the stoichiometry of marine  
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26 phytoplankton.  
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32 Primary production in most modern aquatic environments is limited by either N or P. Large  
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34 areas of the modern surface ocean appear to be limited by nitrogen (e.g., Falkowski et al.,  
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36 1998). However, it is generally assumed that extended deficits in bioavailable N could be  
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38 mitigated by biological fixation of atmospheric N<sub>2</sub>, given sufficient availability of other  
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40 nutrients. Notably, this scenario assumes a nearly limitless supply of volcanically-derived N<sub>2</sub>  
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42 in the atmosphere, despite recent suggestions that atmospheric N<sub>2</sub> could have varied over  
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44 Earth history, following changes in N burial rates and subduction over Earth history (as  
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46 discussed below). Phosphorus is primarily sourced from continental weathering, without a  
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48 significant atmospheric source. This dependence on riverine delivery, which is tied to  
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50 tectonic processes via uplift and erosion, suggests that P is the limiting nutrient on timescales  
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52 of greater than thousands of years (Tyrrell, 1999). In addition, the importance of Fe and other  
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54 bioactive metals as micro-nutrients has become increasingly apparent in recent decades (e.g.,  
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56 Frausto da Silva & Williams, 2001; Martin, 1990). Changes in the sources and sinks of these  
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3 redox-sensitive metals have been mainly tied to oxidative weathering and fluctuations in  
4 ocean redox chemistry, particularly during the Precambrian (e.g., [Anbar & Knoll, 2002](#)).

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8 Tectonic processes can therefore exert a significant control on Earth surface redox and  
9 climate through their involvement in global nutrient cycling and primary productivity. These  
10 “biogeodynamic” controls fall under two categories: 1) A *direct control* via subduction and  
11 volcanism (e.g., for CO<sub>2</sub> and N<sub>2</sub>); or, 2) An *indirect control* via burial, uplift, and erosion  
12 (e.g., for phosphorus and some bioactive trace elements). Here I examine these connections  
13 within the context of global nutrient cycling, and their possible implications for Earth system  
14 evolution over geologic history.  
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### 25 **Biogeodynamic Controls on Atmospheric N<sub>2</sub>**

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27 Dinitrogen gas constitutes 78% of the modern atmosphere. As the dominant atmospheric gas  
28 on Earth, the partial pressure of N<sub>2</sub> has an important effect on surface temperature and  
29 planetary habitability ([Goldblatt et al., 2009](#); [Stüeken et al., 2016](#); [Zahnle & Buick, 2016](#)).  
30  
31 Despite this significance, the history of  $pN_2$  over geologic time is not well understood, and  
32 empirical constraints on past atmospheric pressure are limited. N<sub>2</sub>/<sup>36</sup>Ar systematics in fluid  
33 inclusions ([Nishizawa et al., 2007](#); [Marty et al., 2013](#); [Avicé et al., 2018](#)), fossilized raindrop  
34 imprints ([Som et al., 2012](#); [Kavanagh & Goldblatt, 2015](#)), and gas bubbles in basaltic lavas  
35 formed at sea level ([Som et al., 2016](#)) have all been used to estimate  $pN_2$  at  $\geq 2.7$  Ga ([Figure](#)  
36 [1A](#)), but a consensus is lacking. Nitrogen flux estimates from modern arc systems have  
37 further complicated the picture, with some calculations suggesting a net outflux of nitrogen  
38 from the mantle or zero sum game (e.g., [Fischer et al., 2002](#)), while others suggest a net  
39 influx of nitrogen from Earth’s surface to the mantle (e.g., [Barry & Hilton, 2016](#); [Busigny et](#)  
40 [al., 2011](#); [Goldblatt et al., 2009](#); [Mallik et al., 2018](#)).  
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3 Determining the history of atmospheric  $pN_2$  over time requires a detailed understanding of  
4 the processes acting to cycle nitrogen from Earth's surface to the mantle, and vice versa. The  
5 modern surficial N cycle is largely driven by biological processes. The great majority of  
6 biologically available N in the ocean-biosphere system comes from nitrogen fixation, the  
7 biological uptake of atmospheric  $N_2$  into biomass ( $N_{org}$ ) (Figure 2). Early in Earth's history,  
8 abiotic  $N_2$  fixation reactions induced by lightning or photochemical processes could have  
9 provided small amounts of nitrogen to the biosphere (Navarro-Gonzalez et al., 2001).

10 Hydrothermal fluids can also recycle sedimentary ammonium back into the water column at  
11 mid-ocean ridges and associated hydrothermal vent systems (Lilley et al., 1993) providing a  
12 secondary source of dissolved inorganic nitrogen (DIN). Weathering of N-rich continental  
13 rocks has additionally been proposed as a significant contribution to the terrestrial N cycle  
14 (Houlton et al., 2018), and possibly to the atmosphere (Stüeken et al., 2016). However, these  
15 inorganic sources of DIN (shown in red in Figure 2) would have constituted only a small  
16 percentage of the global supply of fixed N to surface oceans once  $N_2$  fixation evolved very  
17 early in Earth history (Boyd & Peters, 2013; Stüeken et al., 2015; Weiss et al., 2016).

18 Organic N is released as ammonium ( $NH_4^+$ ) during degradation of biomass, which is quickly  
19 recycled and re-assimilated. In the presence of oxygen, ammonium can also be oxidized to  
20 nitrite and nitrate, which provide an additional source of dissolved inorganic nitrogen (DIN)  
21 to the biosphere. A significant amount of this DIN is returned to the atmosphere via anaerobic  
22 processes in oxygen-minimum zones (e.g., Dalsgaard et al., 2005). Over geologic timescales,  
23 small amounts of fixed nitrogen can escape biological recycling and be buried in the  
24 sediments (Schroeder & McLain, 1998). During burial, nitrogen-bearing rocks can undergo  
25 metamorphism, which returns a significant fraction of the nitrogen back to the atmosphere  
26 (e.g., Haendel et al., 1986). The remainder of sedimentary N is subducted, along with  
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3 ammonium from altered oceanic lithosphere (Busigny et al., 2011; Halama et al., 2014),  
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5 forming the primary flux of N from the surface to the deep.  
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8 Once subducted, the fate of N in the mantle is less clear. Recent experimental and theoretical  
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10 constraints suggest that the speciation of nitrogen is highly variable under a range of mantle  
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12 redox and pH conditions, e.g., with ammonium dominating over molecular N<sub>2</sub> under more  
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14 reducing mantle conditions (Li & Keppler, 2014; Mikhail et al., 2017; Mikhail & Sverjensky,  
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16 2014). Nitrogen speciation in the mantle is critical to the volcanic resupply of N<sub>2</sub> to the  
17  
18 surface environment, because molecular N<sub>2</sub> is likely to be out-gassed, while ammonium could  
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20 be stored in mineral phases in the deep Earth (Mikhail & Sverjensky, 2014). These  
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22 constraints indicate that the storage capacity for N in the mantle could be significantly larger  
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24 than previously recognized (e.g., Cartigny & Marty, 2013; Johnson & Goldblatt, 2015). For  
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26 example, recent estimates suggest that silicate minerals in Earth's interior could store up to  
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28 50 times present atmospheric levels of nitrogen (Yoshioka et al., 2018).  
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34 The capacity for Earth's mantle to store large reservoirs of N in the form of ammonium-  
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36 bearing silicates could provide a mechanism for changing atmospheric  $pN_2$  over geologic  
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38 timescales. As above, the drawdown of N<sub>2</sub> from the atmosphere into the sediments is  
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40 fundamentally controlled by the operation of the biogeochemical N cycle; once subducted,  
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42 the fate of sedimentary and igneous N depends on conditions in the mantle wedge, including  
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44 the redox state of associated fluids. Both of these parameters have evolved over geologic  
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46 time, alongside chemical and biological evolution at Earth's surface. Zerkle and Mikhail  
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48 (2017) estimated that the increasing efficiency of the global biosphere could have driven the  
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50 burial of organic N up to 50% of modern levels by the early Proterozoic (Figure 1B).  
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55 Notably, these estimates assumed a conservative burial efficiency of 10%, similar to the  
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57 modern, but burial rates could have been even higher under the anoxic depositional  
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59 conditions of the Precambrian deep oceans (e.g., Kipp & Stüeken, 2017). Recent models of  
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3 atmospheric N<sub>2</sub> dynamics suggest that large changes in N<sub>2</sub> fixation into biomass coupled with  
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5 significantly higher overall rates of biomass burial could drawdown  $pN_2$  to values less than  
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7 0.5 bar by 2.7 Ga (Stüeken et al., 2016).  
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11 In addition, changes in the speciation and degassing of N in Earth's mantle associated with  
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13 changes in Earth surface oxygenation could have contributed to large swings in atmospheric  
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15  $pN_2$ . A significant amount of subducted rock N could have been stored as ammonium  
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17 throughout the Precambrian, given that deep oceans remained largely anoxic and ferruginous  
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19 until the Neoproterozoic, and perhaps even into the early Phanerozoic (Stolper & Keller,  
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21 2018). Considering only the available constraints on changes in N speciation with redox  
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23 (Mikhail & Sverjensky, 2014), if the mantle wedge remained largely reducing, a net influx of  
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25 N<sub>2</sub> from the atmosphere-biosphere system to the mantle could have been the norm for much  
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27 of Earth history. Thus, N<sub>2</sub> levels could have started out higher than today (e.g., Goldblatt et  
28  
29 al., 2009), and been pulled down to modern levels or even lower (as the illustrated by the  
30  
31 upper bounds in the "Flux inversion model" in Figure 1A). In contrast, N<sub>2</sub> drawdown from  
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33 near modern levels could have led to lower  $pN_2$  in the Archean (as shown by the lower  
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35 bounds of the "Flux inversion model") (e.g., Som et al., 2016). Following the Great  
36  
37 Oxidation Event in the Paleoproterozoic, widespread O<sub>2</sub>-dependent nitrogen loss from the  
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39 biosphere back to the atmosphere (e.g., Zerkle et al., 2017), coupled with an enhanced N<sub>2</sub>  
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41 release during oxidative weathering (Stüeken et al., 2016), could have somewhat eased (or at  
42  
43 least slowed) N<sub>2</sub> drawdown. Once deep waters became oxygenated in the early Phanerozoic,  
44  
45 subduction zones injected oxidizing fluids into the mantle wedge, oxidizing NH<sub>4</sub><sup>+</sup> to N<sub>2</sub>, and  
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47 promoting outgassing to modern  $pN_2$  levels.  
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55 These theoretical scenarios for  $pN_2$  changes through time are difficult to test given available  
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57 data. In a pioneering study, Johnson and Goldblatt (Johnson & Goldblatt, 2017) provided  
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59 evidence for enhanced incorporation N into continental rocks over geologic time, in the form  
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3 of glacial tills and granites (Figure 1B). Additionally, the  $\delta^{15}\text{N}$  ratios of these rock-bound  
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5 nitrogen species support a biological origin. Intriguingly, the distribution of N in glacial tills  
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7 also seems to vary with paleogeography, with total N abundance in Asian tills  $\gg$  South  
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9 American and African tills  $\gg$  North American tills. This geographical disparity suggests  
10  
11 some local controls on N drawdown, from variable continental growth histories, changes in  
12  
13 local biogeochemical N cycling, and/or changes in subduction zone conditions (Johnson &  
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15 Goldblatt, 2017). This sparse dataset offers a tantalizing picture of changes in the  
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17 geobiological N cycle over Earth history, but additional geochemical and geodynamic data is  
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19 required to tease out the mechanism(s) behind these spatial and temporal heterogeneities.  
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### 24 **Phosphorus Burial and Continental Weathering**

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26 Phosphorus is generally believed to be the limiting nutrient for global primary productivity  
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28 on  $\geq 200$  ka timescales (Tyrrell, 1999) because there is no limitless supply of phosphorus to  
29  
30 the oceans. Instead, the input of P to the marine biosphere relies on riverine delivery from  
31  
32 land. Once delivered to the oceans, P exists mainly as  $\text{P}_{\text{org}}$  and as P adsorbed to Fe-  
33  
34 (oxyhydr)oxides (hereafter referred to as “Fe-oxides” for simplicity). As with nitrogen, the  
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36 burial of phosphorus in deep sea sediments and subsequent subduction constitute the main  
37  
38 flux of P to the mantle. However, P undergoes complex diagenetic cycling in marine  
39  
40 sediments, and the extent to which P is trapped versus recycled into the water column is  
41  
42 highly redox dependent. Under a well-oxygenated water column, phosphorus can be released  
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44 during respiration of organic matter or reduction of Fe-oxides, but the released P is either  
45  
46 efficiently scavenged by re-adsorption to Fe-oxides near the sediment-water interface, or  
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48 forms authigenic carbonate fluorapatite (CFA) (e.g., Ruttenberg & Berner, 1993; Slomp &  
49  
50 van Raaphorst, 1993). Under anoxic conditions, the Fe-oxide trap at the sediment surface is  
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52 diminished, and more P is released into the water column (e.g., Ingall et al., 1993; Dellwig et  
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3 [al., 2010](#)), particularly under euxinic (anoxic and sulfidic) depositional conditions when Fe-  
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5 oxides are reduced by dissolved sulfide ([Poulton, 2003](#)).  
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9 The dependence of P cycling on oxygen levels implies that spatial and temporal changes in  
10  
11 ocean redox chemistry over geologic time would also have controlled the amount of P that  
12  
13 was buried in weatherable continental sediments and deep-sea oceanic sediments prior to  
14  
15 subduction. In the Fe-rich oceans of the Archean, P burial rates could have been universally  
16  
17 high due to extensive sorption onto Fe-oxides (e.g., [Bjerrum & Canfield, 2002](#); [Jones et al.,](#)  
18  
19 [2015](#)) and a paucity of electron donors for organic matter remineralization ([Kipp & Stüeken,](#)  
20  
21 [2017](#)). The redox-stratified oceans that characterized the Proterozoic ([Poulton & Canfield,](#)  
22  
23 [2011](#)) could have promoted high P burial in ferruginous deep-water sediments ([Reinhard et](#)  
24  
25 [al., 2017](#); cf, [Planavsky et al., 2010](#)), but significant recycling of P back to the water column  
26  
27 at widespread sulfidic continental shelves ([Poulton, 2017](#)). Recent studies also suggest that  
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29 Fe(II) phosphate (vivianite) could be an important sink for reactive P during early diagenesis  
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31 in low-sulfide sediments or microenvironments (e.g., [Jilbert & Slomp, 2013](#); [Dijkstra et al.,](#)  
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33 [2016](#)), and vivianite has been proposed as a further important sink for phosphate in the mid-  
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35 Proterozoic ([Derry, 2015](#)).  
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42 In addition to the burial-subduction controls on P cycling to the deep Earth, the flux of P to  
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44 the biosphere relies on the weathering of P-bearing minerals, which is indirectly linked to  
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46 tectonics via uplift and erosion. A weathering driver for P delivery relies on the availability of  
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48 continental freeboard, although the net growth rate for continental crust on the early Earth is  
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50 controversial (e.g., see review in [Dhuime et al., 2017](#)). For example, most estimates suggest  
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52 continental freeboard could have been scarce prior to ~3.0 Ga (e.g., [Cawood et al., 2013](#);  
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54 [Pons et al., 2013](#)); alternatively, net growth of continental crust could have been complete by  
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56 the end of the Hadean (e.g., [Armstrong, 1981](#); [Rosas & Koranaga, 2018](#)). Once large  
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58 continental masses appeared, enhanced P weathering could be linked to the supercontinent  
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3 cycle, either by the redistribution of crustal mass to higher relief during supercontinent  
4 assembly or by the formation of continental margins during supercontinent breakup.  
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8 An enhanced flux of phosphate to the biosphere associated with tectonic activity and several  
9 extensive glaciations has been implicated in organic carbon burial and global oxygenation in  
10 the late Paleoproterozoic (e.g., [Bekker & Holland, 2012](#); [Papineau et al., 2013](#)). Two mantle  
11 superplume events (one from ~2.45-2.48 Ga and one at around 2.2 Ga) initiated continental  
12 rifting and led to the eventual breakup of the Kenorland supercontinent at ~2.1 to 2.0 Ga  
13 ([Aspler & Chiarenzelli, 1998](#); [Barley et al., 2005](#); [Halls et al., 2008](#); [Heaman, 1997](#)). The  
14 formation of epicratonic rift basins during the breakup of Kenorland could have created  
15 continental configurations that promoted the upwelling of deep P-rich waters, stimulating  
16 primary production. Increased chemical weathering fluxes of phosphorus to the oceans after  
17 major glaciations could also stimulate photosynthetic oxygen production ([Lenton & Watson,](#)  
18 [2004](#)). In addition, generation of H<sub>2</sub>SO<sub>4</sub> (sulfuric acid) during pyrite oxidation on the  
19 continents could have decreased the pH of soils and groundwaters (e.g., [Holland, 2002](#);  
20 [Konhauser et al., 2011](#)), further promoting apatite dissolution on land and bioavailable P  
21 delivery to the oceans ([Bekker & Holland, 2012](#); [Guidry & Mackenzie, 2003](#)).  
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41 Similarly, P fertilization from the weathering of large igneous provinces has been suggested  
42 to have contributed to the Neoproterozoic oxygenation event and ensuing radiation of  
43 macroscopic life ([Planavsky et al., 2010](#); [Horton, 2015](#); [Laakso & Schrag, 2017](#)). Biological  
44 colonization of the land surface could have further amplified mid- to late- Neoproterozoic  
45 weathering rates via selective weathering of P from rocks with organic acids ([Lenton &](#)  
46 [Watson, 2004](#)). As above, enhanced phosphorus scavenging in anoxic, Fe-rich oceans  
47 ([Reinhard et al., 2017](#)), coupled with global tectonic quiescence ([Cawood & Hawkesworth,](#)  
48 [2014](#)) could also have helped to maintain the low oxygen levels that typified much of the  
49 “boring billion” years in between.  
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3 While these and other feedbacks between planetary redox, P availability and burial have  
4 received a significant amount of attention (e.g., Bekker et al., 2003; März et al., 2008), the  
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6 inferred connection with tectonic processes is not well established. Cox et al. recently  
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8 examined the phosphorus content of igneous rocks as a proxy for the continental P inventory  
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10 (Cox et al., 2016; 2018). Notably, the average of the measured igneous  $P_2O_5$  values are  
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12 similar to  $P_2O_5$  values measured in glacial tills (although slightly higher) (Gaschnig et al.,  
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14 2016), and both show a general trend of increasing  $P_2O_5$  through time (Figure 3). Cox et al.  
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16 interpreted this trend to reflect changes in continental P content due to secular changes in  
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18 mantle cooling, and linked this to atmospheric oxygenation. These researchers also identified  
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20 secondary variations in crustal  $P_2O_5$  content associated with mafic crustal production, and  
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22 attributed these differences to the behavior of P during melting and fractional crystallization.  
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24 This study provides a critical first step in identifying the behavior of P in igneous systems,  
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26 and provides a number of interesting, and testable, hypotheses. If we further examine their  
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28 dataset in the context of the supercontinent cycle (Figure 3), there also appear to be peaks in  
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30  $P_2O_5$  roughly associated with continental breakup, at least within the given time constraints  
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32 we have on the known supercontinent cycles. Notably, no systematic geochemical studies  
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34 have thus far been conducted to track changes in P cycling and weathering fluxes over a  
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36 supercontinental cycle, likely because of the lack of a straightforward method for estimating  
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38 ancient P concentrations in seawater from the sedimentary rock record.  
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### 49 **Conclusions and Ways Forward**

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51 The above discussion highlights some of the fundamental connections between the surface  
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53 and deep Earth that contribute to global elemental cycling over geologic timescales.  
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55 Sediments subducted into the mantle are highly influenced by biological activity, and their  
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57 composition and abundance will vary following the evolution of life and biogeochemical  
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3 cycles over Earth history. Similarly, the local environment of the mantle wedge can directly  
4 control the speciation and volatility of elements that are subducted within these sediments,  
5 with critical consequences for their storage or release from the mantle into the atmosphere.  
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7 These connections extend well beyond the nitrogen and phosphorus cycles, for example, to  
8 implications for changes in the burial ratio of organic carbon to carbonate-rich sediments or  
9 in the burial of sulfur as primarily sulfide to sulfate. However, we currently lack fundamental  
10 constraints on the relevant reservoirs, in particular the speciation and the residence times of  
11 elements in the deep Earth. Similarly, uplift and erosion of continental rocks provide an  
12 important supply of nutrients to the global biosphere, but their temporal distribution and  
13 availability to weathering reactions will be dependent on their behavior and resulting  
14 mineralogy during different crustal production mechanisms.  
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29 Understanding how these mechanisms have acted in time and space has generally been a  
30 problem for geodynamicists; however, the implications for the biosphere make this of  
31 fundamental importance to biogeochemists as well. Quantifying these connections should be  
32 an important target for Earth scientists across disciplines, with priority research directions  
33 including but not limited to:  
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- 41 1) Experimental and theoretical examinations of the speciation and fate of bioactive  
42 elements during subduction under varying mantle conditions, along with implications  
43 for the volatilization and return of these elements to Earth's surface;  
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- 48 2) Development of novel tools and proxies for assessing atmospheric pressure from the  
49 rock record, and application of these (and already established) tools to more recent  
50 time periods in Earth history (particularly the mid-Proterozoic);  
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- 55 3) A systematic study of the distribution and weathering susceptibility of P- and  
56 micronutrient-bearing minerals under different tectonic regimes;  
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- 4) The generation of high-resolution datasets for crustal P and micronutrients across a supercontinental cycle, preferably coupled directly to weathering proxies, such as Li or Mg isotopes (e.g., [Pistiner & Henderson, 2003](#); [Teng et al., 2010](#)); and, finally,
- 5) The incorporation of deep Earth reservoirs and elemental speciation in Earth system models of global nutrient cycling (e.g., [Mills et al., 2014](#)), and, as a corollary, the inclusion of geochemical and biologically-mediated sediments in geodynamic models of subduction and volatilization.

These target research areas and similar ways of thinking will allow us to gain a more holistic view of Earth system evolution over geologic timescales, and to determine the impact of biogeodynamic connections on climate, global redox, and ultimately life.

#### **Data accessibility**

All data utilized in this manuscript are available in the cited references.

#### **Competing interests**

The author declares no competing interests.

#### **Acknowledgements**

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#### **Figure captions**

Figure 1. A. Empirical constraints on  $pN_2$  over Earth history. Constraints based on fluid inclusion data (in blue) are from [Marty et al. \(2013\)](#) and [Nishizawa et al. \(2017\)](#); fossil raindrop imprints (in green) are from [Som et al. \(2012\)](#) and [Kavanaugh and Goldblatt \(2015\)](#); gas bubbles (in red) are from [Som et al. \(2016\)](#). Also shown is a hypothetical scenario for the evolution of atmospheric  $pN_2$  (in brown), based on [Zerkle & Mikhail \(2017\)](#), as described in the text. B. Estimates of nitrogen burial through time (based on [Zerkle & Mikhail, 2017](#)),

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3 along with the measured N content of glacial tills and granites (from [Johnson & Goldblatt,](#)  
4 [2017](#)).

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8 Figure 2. The combined surface and deep Earth nitrogen cycle. Abiotic fluxes are shown in  
9 red, biological fluxes are shown in blue.  $N_{\text{org}}$  = organic nitrogen;  $N_{\text{sed}}$  = N incorporated into  
10 sediments; DIN = dissolved inorganic nitrogen.

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16 Figure 3. Total  $P_2O_5$  measured in glacial tills (purple diamonds; from [Gaschnig et al., 2016](#))  
17 and in igneous rocks, including the averages (blue circles) and 95% confidence intervals  
18 (blue field; from [Cox et al., 2018](#)), compared to established supercontinent cycles.

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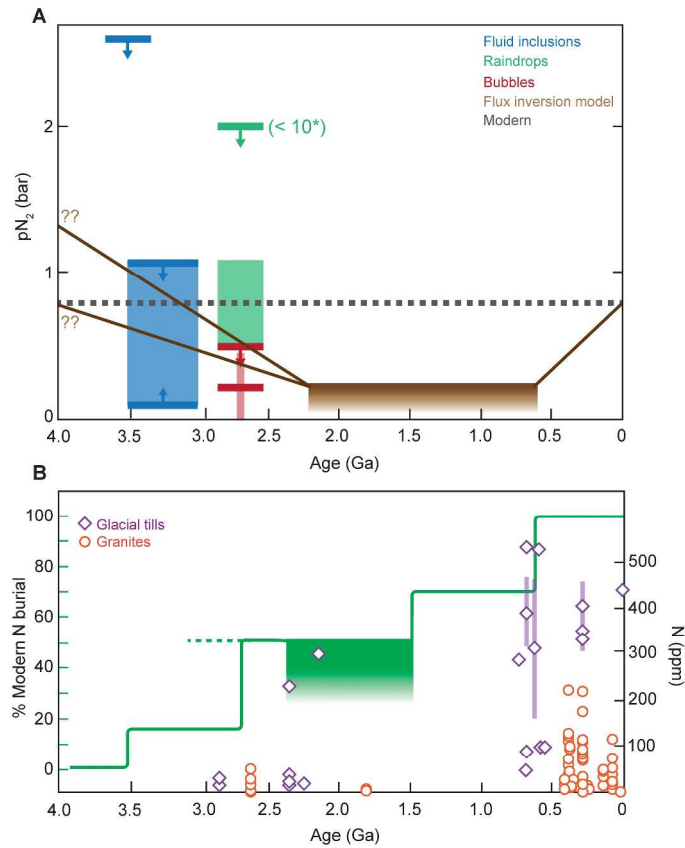


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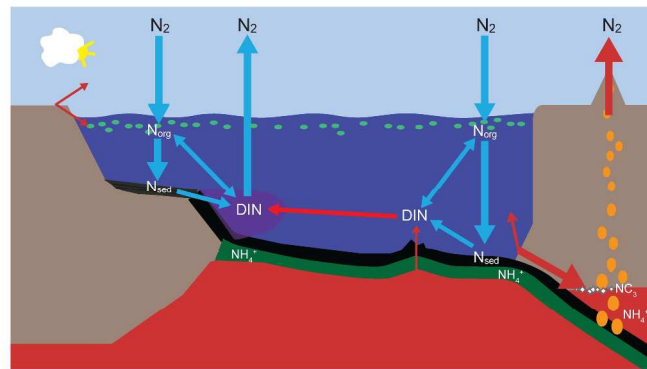
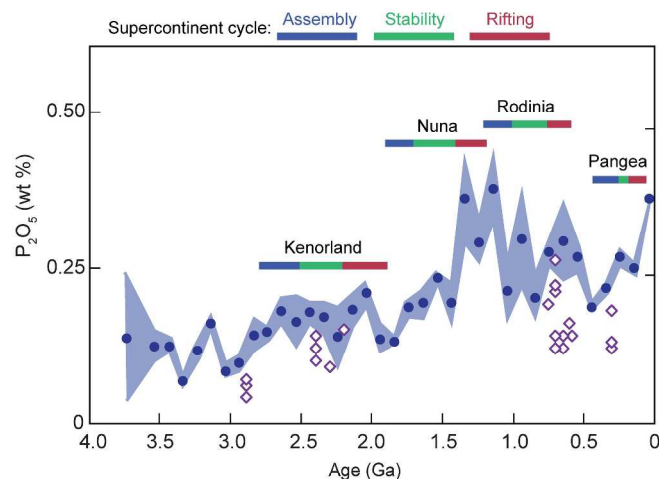


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Figure 3. Total  $P_2O_5$  measured in glacial tills (purple diamonds; from Gaschnig et al., 2016) and in igneous rocks, including the averages (blue circles) and 95% confidence intervals (blue field; from Cox et al., 2018), compared to established supercontinent cycles.

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297x420mm (300 x 300 DPI)