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Review

Life's Critical Role in the Long-term Carbon Cycle: the Biotic Enhancement of Weathering

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Abstract: The biotic enhancement of weathering (BEW) has important implications for the long-term carbon cycle, in particular as a driver of climatic cooling. The BEW factor is defined as how much faster the silicate weathering carbon sink is under biotic conditions than under abiotic conditions at the same atmospheric pCO_2 level and surface temperature. The BEW factor and its evolution over geological time can potentially be inferred from consideration of empirical and theoretical weathering studies. Estimates of the global magnitude of the BEW are presented, drawing from laboratory, field, watershed data and models of the long-term carbon cycle, with values ranging from one to two orders of magnitude.

Keywords: biotic enhancement of weathering; long-term carbon cycle; chemical; physical and biological weathering

1. Introduction

Understanding processes in the critical zone, the site of rock weathering for the subaerial crust, is a wide ranging international research program [1]; e.g., [2–6]. In this contribution I will make the case for inferring an estimate of the magnitude of the biotic enhancement of weathering (BEW) factor and its evolution over geological time from a consideration of empirical and theoretical weathering studies. Estimates of the global magnitude of the BEW can be derived from laboratory, field, watershed data and models of the long-term carbon cycle. The BEW factor is defined as how much faster the silicate weathering carbon sink is under biotic conditions than under abiotic conditions at the same atmospheric pCO_2 level and surface temperature [6]. I have previously argued that the BEW factor and its evolution over geologic time has had a critical role as a driver of climatic cooling, since it makes possible the same carbon sink flux as for the case of an abiotic regime at steady-state equal to the volcanic source, but at a lower climatic temperature and atmospheric pCO_2 level [6,7]. If this BEW factor is significantly greater than one, on an abiotic Earth this steady-state would occur with higher atmospheric carbon dioxide levels and surface temperatures generating an equal flux as the outgassing flux, than the biota on land and in its subsurface does at lower atmospheric pCO_2 levels and temperatures by the multifold processes included in BEW. Given my earlier assessment of the present global BEW on the two orders of magnitude (7), we will now revisit this estimate in light of the extensive research on weathering and the long-term carbon cycle since then.

The likely contributors to the present BEW with forest and grassland ecosystems/soils as the main sites include [8,9], with more recent papers noted:

1. Soil stabilization with its contribution to reactive mineral surface, while Earth surface close to abiotic has little or no soil or regolith, so the land surface reacting with water/carbon dioxide is closer to two-dimensional, with a lower surface roughness, noting the presence of cracking/porosity in the upper level of the crust.

2. pCO_2 elevation/carbonic acid generation in soil from root/microbial respiration and decay of organic matter driving the chemical weathering of CaMg silicates [10], the basis for their concept of BEW; reference [11] argues that this process driven by ecosystem metabolism plays an important role in mineral weathering.

3. Organic acids/chelators in soil, the multifold interactions in the rhizosphere/mycorrhizae including biogeophysical/biochemical weathering by soil fungi driving the breakup and digestion of CaMg silicate mineral particles (e.g., see [12–18]. Biological weathering resulting from mycorrhizal evolution include the following mechanisms: acidification due to hydrogen ion and organic exudates as well as the result of respiration and elevated $pCO_2(g)$, litter decomposition and carbon transfer to heterotrophs and evapotranspiration and stabilization of soil [19];

4. Evapotranspiration contribution to maintaining soil water and runoff (plant biological pump [20];

5. Diffusion of oxygen into soil contributing to oxidation of CaMgFe silicates [21,22];

6. Water retention by soil organics, without water, no chemical weathering;

7. Turnover of soil by ants, earthworms, mixing organic and mineral particles (e.g., Darwin's pioneering research), noting the recent research pointing to the role of ants as strong agents of chemical weathering of silicates [23];

8. Plant root activity in physical weathering, breaking up rock substrate. On bedrock slopes, the cycle of plant colonization, erosion, recolonization facilitates chemical weathering as fresh rock is exposed;

9. Biotic sink effect for mineral nutrients (assuming steady-state soil/biomass, with net flux of Ca, bicarbonate out), or increasing plant biomass storing Ca to be released with organic decay.

There are also biotic effects that act to slow down weathering. The accumulation of thick, depleted soils in low slope terrains act as a barrier to water flow to fresh bedrock Old tropical soils can be severely depleted in nutrient elements [24], implying a slow-down in chemical weathering of the rock substrate. However, in tropical ecosystems, erosion on hillslopes can act to increase chemical weathering [25]. While the forest canopy slows down raindrop erosion, there is evidence of significant bedrock-derived chemical weathering in lowland humid tropical weathering, plausibly enhanced by the biota, in particular by deep root systems and their associated mycorrhizae compared to abiotic exposed bedrock [26–28]. Macropores created by earthworms, ants and other animals in soils allow water flow bypassing saturated pore space, hence no active chemical weathering of silicate minerals in the soil, but on the other hand water flow may as a consequence be diverted to fresh bedrock promoting chemical weathering.

Lichen coverage of bedrock in most cases promotes chemical denudation, so it is not an example of biotic retardation of weathering (BRW). But even with these retarding effects, BEW dominates globally. This conclusion will be demonstrated by the BEW estimates in what follows and their implications to a global scale with its heterogeneous weathering regimes. In other words, BRW processes simply subtract from the overall effect of processes which enhance the weathering rate over the abiotic at the same atmospheric pCO_2 and surface temperature level.

A carbon dioxide source to the atmosphere results from the reaction of calcium carbonates with sulfuric acid produced by the oxidation of sulfide minerals such as pyrite, with the ratio of this carbon dioxide source to the silicate weathering sink increasing with erosion driven by tectonic uplift [28–30]. Since atmospheric oxygen is a biological product, the net impact of sulfide oxidation is an effective subtraction from BEW as defined. The importance of this process to the long-term carbon cycle is still under debate. Apparently, the variation in the strength of the silicate weathering feedback is sufficient to understand mass balance of the carbon cycle in the Cenozoic [31], while sulfide oxidation source prevented runaway glaciation in the Quaternary [32].

Here is a corollary hypotheses relating weathering, erosion and biology: biology-driven weathering dominates the weathering sink of carbon in the long-term carbon cycle. In eroding landscapes, weathering front advance at depth is coupled to surface denudation via biotic processes (hypothesis 4 [1]).

2. Discussion

2.1. The increase of BEW over geologic time

The evolution of and progressive increase in BEW over geologic time is summarized in Table 1.

Table 1. The Coevolution of Biota and the Biosphere, Resulting in the Increase of BEW over Geologic Time.

Hadean/early Archean
Prokaryotic soil crusts, invasion into critical zone (methanogens, anoxygenic phototrophs, early heterotrophs); soil stabilization and microbial dissolution
Archean
2.8 Ga Cyanobacteria (boost in productivity on land, oxygenic photosynthesis)
2.4 Ga Rise of atmospheric oxygen; oxidation in critical zone weathering (e.g., promoting spheroidal weathering, microfracturing, [21,99,100]
Late Archean, Proterozoic
Ice wedging, promoting physical weathering, erosion; eukaryote/prokaryote symbioses on land (increase in biotic productivity).
Late Proterozoic, Phanerozoic
Emergence of Fungi (lichens), Bryophytes, Vascular Plants, Forests
Rhizosphere and mycorrhizal evolution, first arbuscular mycorrhizal fungi, followed by ectomycorrhizal fungi in Cenozoic [13,15,18,19]
Further growth in biotic productivity on land drives higher critical zone pCO_2 levels and amplifies the plant biological pump [20]

The earliest biota on land, primitive prokaryotes had modest enhancement of weathering, with the coevolution of biota and the biosphere driving the increase in terrestrial biotic productivity and capacity to sequester nutrient elements from the crust. Prior to the rise of atmospheric oxygen at about 2.4 Ga life on the land and in shallow water environments would be challenged by the high ultraviolet flux, so it is plausible that protective mechanisms emerged in the Hadean/Archean, including cellular coating with secondary mineral formation (117–121). The emergence of ice wedging in the Late Archean/Proterozoic is postulated as a result of long-term surface cooling from a hot-house to an ice-house climate over geologic time [7]. Thus, ice wedging is included as a BEW process because of BEW-promoted climate cooling reached the water freezing limit on mountains, *not because a role of biology is claimed to be necessary to induce ice wedging*. In the late Proterozoic to the present the co-evolution of the biosphere and biology resulted in an additional step-like increase in BEW. As commonly defined, the biosphere consists of the biota and the environment it inhabits and directly influences, i.e., the upper crust, surface hydrosphere and atmosphere. For example, assuming a very warm climate in the Archean to be discussed in 2.3, with the climatic temperature being a constraint of biologic evolution, a cooling climate acts as a prime releaser for new emerging biology, e.g., phototrophs, followed by eukaryotes, metazoa, plants, each with an intrinsic upper temperature limit of growth.

2.2. Field/Experimental (Mesocosm) estimates of BEW

Lichen Weathering

Table 2. Studies measuring Lichen Weathering Flux.

Mini-watersheds in field
(Chemical) weathering enhancement compared to bare rock
Hubbard Brook, New Hampshire, U.S.: Metasedimentary quartz and mica schist
Reference [101]
16 (Mg)
4.4 (Si)
0.09 (Fe)
Wanaque, New Jersey, U.S.: Hornblende Granite
Reference [102]
2.5 (Mg)
3.5 (Ca)
1.9 (Si)
0.5 (Fe)

Table 3. Studies on Lichen-induced Weathering Rinds.

(Chemical) weathering enhancement factor compared to bare rock
Mauna Loa and Kilauea, Hawaii: Basalt flows
Reference [103]
15–71
Hualalai Volcano, Hawaii: Basalt
Reference [104]
2–18
Lanzarote, Canary Islands Basalt
Reference [105] (Source: [102])
16

Lichens are now the dominant vegetation of 8% of the land surface [33], with likely even greater importance in the late Proterozoic/early Phanerozoic prior to the emergence of plants. Lichen weathering offers a more easily constrained system for research on BEW than watersheds and of course the global scale. Lichens enhance chemical weathering of rock substrates by a variety of mechanisms including fungal hyphal digestion of silicate mineral particles [34] as well as promotion of physical breakup of substrate, retention of water, production of carbonic, oxalic acids and chelating agents [35,36]. Tables 2 and 3 summarizes the field and experimental studies of lichen weathering with their estimates of chemical weathering enhancements relevant to BEW.

Plant BEW

Table 4. Mosses (Microcosm experiment).

(Chemical) weathering enhancement factor		
	Ca	Mg
Granite	1.4 ± 0.2	1.5 ± 0.2
Andesite	3.6 ± 0.9	5.4 ± 0.9
Reference [106]		

Table 5. Vascular Plants; Chemical Weathering Studies.

(Chemical) weathering enhancement factor
Skorradulur, Iceland:
Basalt Reference [107]
3–5 (Mg)
2–3 (Ca)
2–3 (Si)
Hubbard Brook, NH:
Granitic outwash (manufactured soil) Reference [108]
18 (Mg)
10 (Ca)

The few studies of bryophyte (moss) weathering and vascular plants are summarized in Tables 4 and 5 respectively. In addition, judging from the results of an experimental study of mineral weathering by a liverwort (with and without a fungal symbiont), a living model of early land plants, early land plants in the early Paleozoic did not apparently significantly enhance chemical weathering over earlier lichen colonization [18]. Rather, the later rise of rooted vascular plants with associated mycorrhizal fungi amplified BEW, especially with the emergence of Devonian forested ecosystems [18]. Recent field research supports this conclusion, with a one-order BEW of vascular plants over "bare" (at least partially lichen and bryophyte-covered) bedrock [37,38]. These results are consistent with the inferred BEW of lichens and plants given previously.

2.3. Inferences from global biogeochemical models

While mesocosm/field studies capture local, likely short-term BEW, global models have the potential to capture long-term global BEW with research results summarized in Table 6.

Table 6. Global Modeling of BEW.

Phanerozoic Modeling
1) Berner lab, culminating in References [109,110]
Inferred Devonian large vascular land plants weathering factor: 4 x
However, more recent papers suggest this estimate is a minimum because atmospheric pCO_2 levels at about 400 mya may have been higher than previously assumed for his weathering factor estimate in 2001 [111], e.g., [112]
2) Beerling lab (From Figure 8, [113]
Inferred Cenozoic Plant Weathering Factor / Berner's plant factor: 2-4 x
(Note as well [13–15,18])
3) Kleidon lab Reference [114]
Removed soil pCO_2 elevation factor from global land Ca flux gives a maximum BEW of 5 x
(Effect limited to regions with active erosion)
Precambrian Modeling
1) Reference [115]
Inferred Phanerozoic BEW: 5.4 x higher than Precambrian
(But suggests progressive growth of BEW in Phanerozoic)
2) Reference [7]
Model calculations assuming a hot Archean climate (2.6 to 3.5 Ga) give an inferred BEW ratio: Present / Archean equal to 25 to 82

Modeling the variation of BEW over Geologic Time

In the following modeling the temperature scenario is assumed to correspond to a hot Archean climate, with global climatic temperatures on the order of 50–70 deg C [7]. My updated case provides a detailed critique of papers which argue for near present temperatures and atmospheric pCO_2 levels in the Archean [39], (e.g., [40–43].

My critique of the widely cited reconstruction of Archean atmospheric pCO_2 levels [40] deserves special attention [39]:

"But isn't the Archean pCO_2 level constrained to be < 0.14 bar by paleosol evidence [44]?

Following the methodology of reference [40, 45], reference [44] base their calculation of inferred atmospheric pCO_2 level on a formula with the formation time of the paleosol in the denominator (e.g., equation (13) in [40]. The critical problem in this approach is the assumption of a similar formation time for the Archean soil as the present, not taking into account the direct temperature effect on mineral dissolution as well as the evidence, both empirical and from modeling, that Archean weathering was much more intense than now. In other words, the conclusion that the pCO_2 level (and temperature) was only modestly higher than today was embedded in their assumptions. Focusing on reference [44] study on a 2.69 Ga paleosol, the derived limit is 10–50 ("best guess" = 41) times the present pCO_2 level. Reference [40] assumes a present climatic temperature to calculate the rate of weathering in its mass balance approach. When a higher temperature (e.g., 60 deg C) at 2.7 Ga is used, the rate of weathering (inversely related to formation time) increases 7 to 23–34 x the present (20 deg C), depending on the assumed temperature (60, 65 deg C; the temperature effect on weathering rate is exponential; assuming an activation energy of 70 kj/mole, and leaving out an additional factor, the runoff as a function of temperature; see discussion of temperature effect on chemical weathering of silicates in reference [7], p.87–89, 150. In addition, another effect of higher temperatures would be to increase the CO_2 gradient and diffusion rate by more rapid reaction with CaMgFe silicates in zone C, the transition to fresh bedrock, thereby shortening the time to produce the paleosol. In other words, a significantly higher assumed temperature at 2.7 Ga would have increased the weathering rate relative to today's climate, and significantly reduced the formation time necessary to create the observed paleosol. Assuming a climatic temperature of 60-65 deg C the computed atmospheric pCO_2 is on the order of 1000 PAL, the product of reference's [40] high limit of 50 x the temperature enhancement factor, with other inputs unchanged, noting that the variation with temperature of the Henry's Law and diffusion constant for CO_2 is much smaller. Hence, the conclusion (low pCO2, low temperature) appears to be embedded in the assumptions used to compute the result." Indeed, a subsequent paper takes into account the impact of temperature on the kinetics of weathering (124). Their modeling for the 2.7–2.8 Ga paleosols includes an upper limit of 40 deg C, constrained by the problematic reference 49. Even at this temperature an upper limit of pCO_2 of 4100 to 13,000 PAL is inferred, well within the range for the high pCO_2 scenario supported here. So aside from the issue of whether the paleosol mineralogy is intact, these results are consistent with a hot Archean climate.

A more recent paper makes a similar case against hot Archean temperatures [46] without presenting any oxygen isotopic support for low temperatures and based on what I take as very problematic arguments, i.e., evidence is presented for Archean gypsum, but this mineral is precipitated under metastable conditions above 70 deg C; see [7], further the presence of "glacial" diamictites is invoked, but the same characteristics are observed in non-glacial diamictites.

New constraints on Archean greenhouse gas levels and climatic temperature ?

A case is made for superflares of the early Sun driving the production of N_2O, a potent greenhouse gas warming the surface of the early Earth, as well as arguing for the destruction of a H_2O/CO_2 atmosphere on early Earth by photodissociation [47]. How extensive this postulated destruction of a H_2O/CO_2 atmosphere is unclear, given the evidence for a Hadean water oceans and high atmospheric pCO_2 levels in the Archean (see evidence presented in [39]). From the measurements of Archean vesicles, it is concluded that the Earth's atmospheric pressure 2.7 billion years ago was less than half of the modern level [48], arguing that "If the mean P_{atm} (0.23 bar) indicated by our data represents the actual atmospheric pressure during the late Archaean, the corresponding boiling point of water at ~58 ^oC would be an upper bound to ambient temperature." An upper limit to P_{atm} of 0.5 bar would of course correspond to a higher boiling point, close to 80 ^oC. This study cites an earlier constraint [49] as supporting evidence, but they fail to mention a critique of the 2012 paper [50] which raised the upper limit to 4.8 times their value. Further, it is not clear whether all the criteria for using basalt vesicles for a pressure estimate were met [51].

New robust support for a hot Archean climate, consistent with the sedimentary oxygen isotopic record, comes from studies on the reconstruction of ancient proteins [52,53,122]. Additional support comes from the measurement of the oxygen isotopic composition of kerogens in Precambrian cherts [54].

While I recognize the case for a hot Archean climate is still under debate, the present BEW ratio to the Archean inferred from its assumption will be discussed in what follows, with conclusions that are consistent with the estimate from current global weathering. This is not a circular argument since the latter estimate is independent from the modeling of the Archean climate.

B_R is defined as equal to B_{Now} / B_t, the ratio of the present BEW to that in the past.

As modeled in [6,7,55] (1),

$$B_R = (A/A_o)(V_o/V) [(P_t/ P_o)^{\alpha} (e^{\beta(Tt-To)}) (e^{\gamma(Tt-To)})] \qquad (1)$$

Where (V/V_o) is the volcanic outgassing flux of CO_2 ratio, (A/A_o) is the land area available for weathering ratio, (P_t/ P_o) is the ratio of atmospheric pCO_2, T_t and T_o are surface temperatures all at time t relative to the present. α, β and γ are constants, whereas the pCO_2 levels were computed from a modified greenhouse function derived from [56] (see [7]). A 3D GCM implies somewhat lower pCO2 levels required for same temperature [57], a conclusion supported by another 3D GCM study [58], however, only temperatures close to the present global average were considered. Assuming this effect persists at temperatures on the order of 50 to 70 deg C, using a 3D greenhouse function would result in modestly lower B_R values than given in Tables 7 and 8, given the smaller influence of pCO_2 level than that of temperature.

Considering the sensitivity of computed B_R values. At time t, the higher the assumed V/V_o ratio, the lower the computed B_R value (greater outgassing drives higher temperatures), and the higher the assumed A/A_o ratio the higher the computed B_R value (greater land area drives lower temperatures).

Table 7. Model results from schwartzman [7].

Parameters for Assumed Variation of Land Area (A) and Carbon Dioxide Outgassing Rate (V) as a Function of Age (t) (after Table 8-3)					
Assumed Model	$(V/V_o) \times (A_o/A)$ at t = 3.8 Ga				
a (constant)	1 x 1 = 1				
b (preferred)	3 x 4 = 12				
c (upper limit)	8 x 10 = 80				
Computed Model Results (after Table 8–4)					
Time	Temperature	Atm. pCO_2	Required B_R		
Ga)	(deg C)	(bars)	a	b	c
3.5	70	2.56	658	82	25
2.6	60	1.22	301	82	43
1.5	50	0.31	103	53	38

Table 8. Model calculations.

Time		B_R	
(Ga)	(A/A_o)*	Model	
		b**	c**
3.5	0-0.2	0-48	0-20
2.6	0.18-0.28	25-40	13-20
1.5	0.39-0.45	26-30	18-20

* Derived from [59]

** Assuming the original V_o/V ratios.

Table 7 shows the model results for the Precambrian from [7] with this methodology being used to recalculate BEW based on the inference of land area from [59] (Table 8). Note that model c in Table 7 is now supported with respect to the ratio of land area by [59] and the ratio of volcanic outgassing rate by reference [60]; also see references in [39].

Revisiting the modeling in light of research regarding the variation of likely land area available for weathering (A/Ao) back in the Precambrian [59].

Assuming constraints on emerged continental area ("A") from [59], taking into account the weatherability factor, higher weatherability is expected in Archean, because of more mafic continental crust and greater area of subaerial oceanic ridges (and oceanic islands of basaltic/komatiitic composition). This effect could be very significant, noting that basalt has a 5–10 times greater weatherability than granite [61,62]. With respect to bicarbonate flux, a basalt/granite ratio of 2 was found, increasing with runoff [63]. Higher Archean land weatherability would increase the computed B_R values, since this would be equivalent to greater effective land area available for chemical weathering under the same climatic temperatures. With the assumptions embedded in this modeling, including hot Archean climates, estimated land areas in the Archean derived from [59] give minimum values of present BEW of 20–48. The agreement of this estimate with studies cited regarding current weathering regimes is conversely support for the hot Archean climate.

Some caveats are in order**.** Global models should take into account other carbon sinks in the long-term carbon cycle, i.e., burial of reduced organic carbon, sea-floor and hydrothermal weathering, in particular their variation over geologic time compared to the critical zone weathering sink. For example, a greater net reduced organic carbon burial now relative to the past would make computed B_R values too high, while a greater relative role of volcanic, sea floor and impact-derived regolith weathering in the Archean would make computed B_R values too low (see [7], again with the assumption of a hot Archean climate scenario. Carbonation of the oceanic crust by reaction with carbon dioxide and water has been proposed as a carbon sink in the early Earth's long term carbon cycle (123). However, the generally higher Archean weathering intensities relative to now implies seafloor weathering was not driving the carbon sink in the long-term carbon cycle, hence a requirement for a hot climate and high atmospheric pCO_2 levels (see references cited in [39].

2.4. Weathering studies at scales from the lab to the watershed; Can the present BEW level be inferred ?

I will now propose an approach to estimating the global BEW from a compilation of weathering studies on basalt on different spatial scales [64], see Figure 1. Given the evidence for a significant artifact effect on dissolution rates derived from ground powders (e.g., [65]) and for parabolically decreasing rates over year timescales in lab experiments [66], I propose using data from a beach sand (not pulverized) experiment on forsterite [67], noting that olivine has the highest dissolution rate of crystalline silicates making up basalt. From reference [67] data, I computed an estimated *geometrically-estimated* surface area and *weathering advance rate of 1 x 10^{-4} mm/year* (plotted on Figure 1). I recognize that using this value is roughly 0.1 the rate derived from a meta-analysis of forsterite dissolution rates at a pH of 5.5 and 25 deg C, from their Fig.13 and 10 [68]. I used this computed rate as an anchor for both field and a proposed abiotic lines to derive an estimated BEW value of 41 at the watershed scale (hot humid climate, data point 9). Note that estimates on the order of 0.1 times the weathering advance rate assumed here for olivine are computed for basaltic volcanic glass, derived from experimental determinations of its roughness factor and dissolution rate (using geometric surface area) at the same conditions give [69,70]. It should be noted that assuming a lower laboratory dissolution rate (e.g., of plagioclase or pyroxene) as an anchor point would result in higher estimated BEW values.

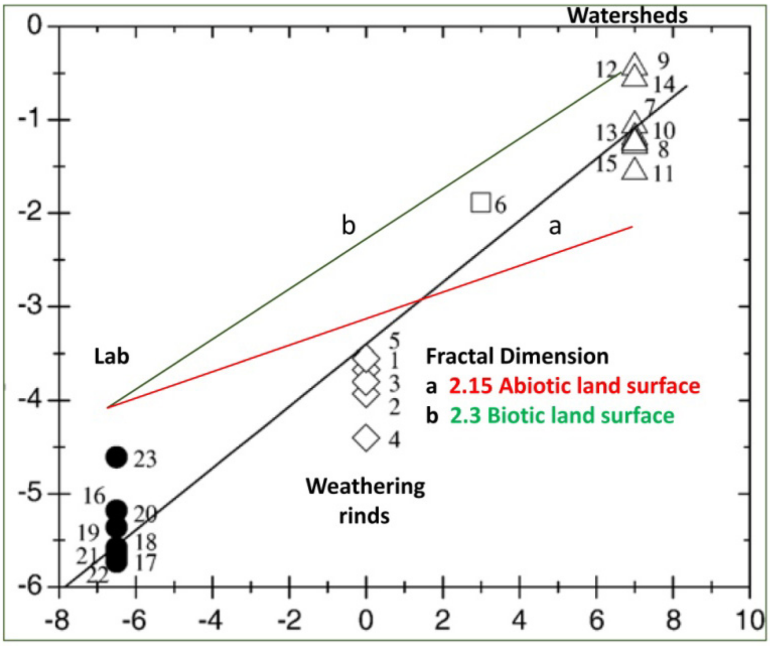

Figure 1. Basalt Weathering (time and scale of measurement)

Y axis: Log (weathering advance rate), in $mm^3\ mm^{-2}\ yr^{-1}$
X axis: Log (scale), in mm
(From Figure 3, Reference [64], lines a and b added)

The abiotic line was generated assuming a roughness fractal dimension of 2.15 corresponding to a sand surface without visible biota [71]. Is this close to a fractal dimension of an abiotic Earth land surface at near modern pCO_2 and temperature levels under consideration? I could not find an estimate of roughness fractal dimensions for bare silicate rock surfaces or high elevation terranes in the literature, but I suspect 2.15 is likely a maximum as a result of wind and water erosion under abiotic conditions. For example, with a value of 2.1, the computed BEW value is 182, a likely maximum.

I submit that the laboratory artifact effect in raising dissolution rates of silicate minerals should be taken more seriously and new experiments with unpulverized samples should be considered, or at least pulverized samples with adequate time provided for relaxation of imposed strain.

Table 9 summarizes limiting values of BEW derived from Figure 1. Note that line b passes through data point 9 which corresponds to a hot humid tropical climate (Java). Therefore this line provides a likely upper limit to BEW for the present global weathering regime. At the watershed level, the difference in denudation rates between lines a and b gives a BEW = $10^{1.61}$ = 41, between b and a line with fractal dimension equal to 2.1 give a BEW = $10^{2.26}$ = 182.

Table 9. Can the Present Global BEW be Inferred?

Line	*Fractal dimension*
a	2.15 (Abiotic land surface ?)
b	2.3 (Earth's land surface)

BEW progressively increases from the laboratory to the watershed

As Figure 1 shows, the weathering advance rate increase with scale, from laboratory experiments to the watershed. The following outlines a case why BEW progressively increases as scale increases. BEW is higher on average for watersheds than for clast weathering because of these factors that kick in at the greater scale:

1) Higher biotic productivity, soil stabilization is generally higher in watersheds, as well as smaller average grain size of CaMg silicates, thereby generating more surface area/volume and greater opportunity for microbial/organic acid interaction. Weathering rinds on clasts are likely closer to 2D systems, analogous to lichen-colonized bedrock exposures.

2) Because of higher biotic productivity in soils underlying forest and grassland ecosystems than more poorly vegetated alluvial fans or glacial outwash sites there should be more water retention and higher pCO_2 levels in watershed soils, as well as higher turnover rates for watershed soils, mixing the organics with the silicates because of greater subsurface biotic productivity.

3) The biotic sink effect should be greater for watershed soils, particularly those forming on slopes, hence a greater flux out of the system of dissolved CaMg and bicarbonates as well as organics.

Note: only the lab data is close to abiotic, with a potential artifact effect from grinding elevating measured dissolution rates.

Biology has increased the surface roughness of the land and the critical zone of weathering by its invasion of the crust and coupling with the atmosphere and climate equivalent to the 5^{th} hypothesis "Biology shapes the topography of the Critical Zone" [1].

There is likely an upper limit to the magnitude of BEW since this factor is potentially self-limiting [72]. Above this upper limit the atmospheric pCO_2 level would plunge below the lower limit potentially for photosynthesis, thereby driving a decline in the biological productivity and global BEW, related to reduced plant and soil activities, with the system being kept at this threshold or going back to higher pCO2 levels, with scenarios dependent on volcanic outgassing and solar inputs. This argument is subject to this caveat: changes in the steady-state of the long-term carbon cycle should take into account the kinetics of flux dependency of oxidation of reduced organic carbon and burial of organic carbon along with a full parameterization of the silicate weathering flux, in addition to the BEW influence at a given temperature and atmospheric pCO_2 level. These considerations hold even if one assumes the volcanic flux and solar luminosity remains constant.

2.5. Some challenges for further research

Global mapping of silicate CO_2 consumption rates can potentially reveal the relative importance of each weathering regime with respect to the global sink flux; where are the hot spots? Tropical volcanic islands have been proposed [73,74], however, there is apparent evidence for a direct magmatic carbon dioxide input to the carbon sink which would reduce the previously estimated silicate CO_2 consumption rates [75].

The cumulative BEW of early land prokaryotes, plus the amplifications from rise of atmospheric oxygen and ice wedging need to be investigated. I propose an experimental approach with mesocosms. Are synergistic interactions important?

The role of BEW in reference's [76] innovative weathering model should be explored, noting their caveat:

"The solute production model does not capture all possible biogeochemical and climatic feedbacks that might operate"

In particular, relevant to this weathering model are the climatic influences on chemical and physical erosion rates, specifically by precipitation affecting local erosion rates via the control on vegetation [77,78]. What are the mechanisms at play for different vegetation types and is there a significant feedback regionally and globally from evapotranspiration? What is the role of fluctuating redox conditions in determining silicate CO_2 consumption rates [79]?

Evidence is presented for a topographic stress control on the generation of bedrock fracturing conducive to chemical weathering [80]. In commentary on [80], reference [81] notes that near-surface cracking can potentially result from ice-wedging and chemical weathering, e.g., driven by by plant roots and the oxidation of ferrous iron-containing minerals such as biotite [82]. How does biology on different scales play a role in determining topography and generation of porosity conducive to chemical weathering in the critical zone?

How do the silicate weathering and reduced organic carbon sinks in the long-term carbon cycle vary with mean global temperature (Tm)? Recent papers point to a weaker temperature dependence than in Berner's BLAG model, mainly because of the critical role of ice-wedging in freeing up exposed rock for weathering, an influence which would be reduced, even eliminated at higher mean global temperatures [83–85].). Reference [83] argues that "At low erosion rates, as temperature and runoff exert progressively less influence on weathering rates [relevant to the long-term carbon cycle], the direct weathering feedback becomes weak to nonexistent." (p. 814). This relationship implies that as Tm increases, the chemical weathering flux will be less than expected from the temperature/runoff dependence on the present Earth surface as well as from simple experimental kinetics because of decreasing erosion rates. Therefore, using a best fit line to determine chemical weathering activation energies with present watershed data should not automatically be extrapolated to higher Tm global climates (e.g., as shown in the temperature dependence of basalt weathering [86]. However, a potential counteracting effect of declining erosion rates with higher

Tm may be the decrease in the ratio of the sulfide oxidation carbon dioxide source to the silicate weathering sink [29,30,87]. As Tm increases, biotic stabilization of soils should persist, though weaken at higher temperatures as the dominant biota shifts from forests/grasslands to microbial cover. In addition, this shift in vegetation type should strengthen the "abiotic" attractor in the bistable plant-soil dynamic [88]. These factors combined with disappearing glacial erosion could potentially weaken the weathering flux dependence on temperature.

Considering the biogeochemical cycle of phosphorus with respect to chemical weathering has promise in better understanding the biotic role in weathering [89,90,74].

Finally, in recent years there has been a major research focus on how deliberate enhancement of chemical weathering could serve as a climate mitigation strategy, with particular attention to using olivine and mafic/ultramafic rocks rich in this mineral as a weathering substrate [91–97]. Research addressing the potential further acceleration of weathering driven by microbiota and plants is even more recent [98]. Given the urgency of the mitigation challenge the potential of BEW acceleration is arguably a fertile area of research.

3. Conclusion

What is magnitude of present BEW?

I conclude that the likely magnitude of BEW factor at present is at the very minimum 10 (more likely closer to 50) to 100. Based on the previously discussed studies, Table 10 shows a proposed evolution to the BEW value represented by modern terrestrial ecosystems. The BEW of the early Proterozoic weathering region corresponds to the combined effects of microbial land colonization/critical zone biota, the rise of atmospheric oxygen and the first appearance of ice wedging in mountains as the climate cooled making this process possible, assuming a hot Archean climate. Their combined value of 5 to 10 is inferred from estimates discussed in this paper.

Table 10. Evolution to the present BEW Factor.

By Early to mid Proterozoic:
Microbial land colonization/critical zone biota + atmospheric oxygen + ice wedging: 5 to 10
Proterozoic to early Paleozoic:
Eucaryotic algae, lichens, bryophyte land biota: 2 to 5, *largely replaced by the*
Early Paleozoic to Present:
Vascular plant ecosystems, culminating in forests and grasslands: 10
Present BEW: roughly 50 to 100

Acknowledgments

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Conflict of interest

The author declares no conflict of interest in this paper.

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