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Wetness controls on global chemical weathering

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Abstract

The formation of soils, the evolution of the biosphere, and the CO2 content in the atmosphere are strongly impacted by chemical weathering. Due to its manifold importance for the long-term stability of the Critical Zone, it is crucial to link weathering rates to the environmental conditions affecting it and develop accurate rate laws for landscape evolution and carbon cycle modeling. Here we use the π theorem of dimensional analysis to provide a theoretical framework to global datasets of weathering rates. As a result, a strong relation between chemical depletion, precipitation and potential evapotranspiration synthesizes the primary role of wetness. Based on this finding, we estimate the spatial distribution of chemical depletion fraction and find that, globally, soils are 50% chemically depleted, 61% of the land is in kinetic-limited conditions, while only 1% is supply-limited. The remaining 38% of the land is in a transitional regime and susceptible to changes in wetness.

1. Introduction

Chemical alteration of the parent material begins when tectonic and topographic stresses fracture the rock and expose minerals to the action of meteoric water (Clair et al 2015, Anderson 2015, Moon et al 2017, Eppes and Keanini 2017, Holbrook et al 2019). From that moment, chemical weathering becomes part of an intertwined system of feedbacks between the Critical Zone and the atmosphere (Gaillardet et al 1999, Kump et al 2000, White and Buss 2014, Pelak et al 2016, Braun et al 2016). As minerals dissolve, nutrients are released, clay minerals precipitate, and soils form, favoring the development of plants and microbial activity (Brady and Weil 2008, Amundson et al 2007, Chorover et al 2011). This in turn increases CO2 production in the upper soil layers (Singh and Gupta 1977, Hanson et al 2000), from which dissolved CO2 is transported to the deeper soil layers (weathering zone) enhancing chemical weathering (Calabrese et al 2017). Weathering reactions then consume the incoming CO2, subtracting it from the atmosphere and mitigating the greenhouse gas effect. An increase in atmospheric temperature and CO2 content in fact would enhance weathering, triggering the negative feedback that stabilizes atmospheric conditions (Berner 1994, Winnick and Maher 2018).

To understand large scale controls on weathering and find empirical laws relating its rate to precipitation, runoff, temperature, and erosion, weathering rates have been measured globally from river chemistry (Gaillardet et al 1999, White and Blum 1995, Dessert et al 2003, Oliva et al 2003, West et al 2005) and from chemical depletion profiles of the regolith (April et al 1986, Brimhall et al 1992, Riebe et al 2003, Rasmussen et al 2011, Dere et al 2013, Ferrier et al 2016, Dixon et al 2016). Although good correlation between weathering rates and these factors is often observed, the lack of consensus on the functional form of the rate laws and the variability in their parameterization (e.g., apparent activation energy) (Dupré et al 2003, Riebe et al 2004, Rasmussen et al 2011, White and Buss 2014, Dixon et al 2016, Maher and Navarre-Stitchler 2019) document the need to disentangle the role of climatic and geologic parameters.

The long list of variables that potentially affect weathering rates has made it difficult to compare measurements from different sites, with sometimes contradictory findings (Gaillardet et al 1999, Dupré et al 2003, West et al 2005,
Dixon et al 2009, Larsen et al 2014, Dixon et al 2016), for instance on the relationship between chemical weathering and erosion. Studies based on reactive transport modeling have proved useful for elucidating the mechanisms driving mineral dissolution at the Darcy scale (Maher 2010, Maher and Chamberlain 2014, Lebedeva and Brantley 2018, Maher and Navarre-Stichter 2019), where the Damköhler number, which compares reaction and fluid timescales, controls the reaction. However, due to the complex heterogeneous structure of the subsurface, idealized conditions (e.g., homogeneous mineral distribution and constant vertical fluid velocity) are typically assumed, so that results remain mostly applicable to a laboratory rather than a field setting, or to a specific site where detailed information on the system is available. Furthermore, the upscaling of the reactive transport equations to soil column or watershed scales remains arduous, due to the mathematical complexity of the system of coupled partial differential equations describing the reactive transport problem (Lasaga 1984, Cederberg et al 1985, Yeh and Tripathi 1991, Saaltink et al 1998, Battiato et al 2009, Battiato and Tartakovsky 2011).

To help resolve the complex interaction between chemical weathering and climate, here we provide a theoretical structure to interpret weathering rate data based on the $\pi$ theorem of dimensional analysis. In complex problems involving a large number of variables, the $\pi$ theorem is often of essential help to derive underlying physical law from experimental data (Barenblatt and Isaakovich 1996). It reduces the number of independent variables by grouping them into dimensionless groups and it allows to obtain scalable laws, which are applicable across sites with very different climatic and geologic contexts. By applying it we show that the chemical depletion fraction—a measure of weathering rate relative to the input of fresh minerals—is related to the ratio of long-term average of potential evapotranspiration over average rainfall (i.e., the dryness index), and not to precipitation or streamflow only. Analysis of chemical depletion data then shows that a strong relation exists between weathering and wetness at the global scale, whereas other factors may dominate local variability. Based on this relation, we estimate the spatial distribution of weathering rates. This allows us to determine the regions of the world under supply or kinetic limitations or more susceptible to wetness. This dimensionless framework can be used to study the relation between weathering rate and all other factors (e.g., surface and subsurface topography) that contribute to the large variability of weathering rates observed globally.

2. The governing dimensionless groups of chemical weathering

We begin by considering the balance equations for mineral and water in the regolith (see appendix A). Tectonic uplift and physical breakdown of the parent material by tectonic and topographic stresses continuously supply fresh minerals, at a rate referred to as denudation rate $D$, which become subject to the action of meteoric water (Anderson 2015). Then, these minerals partly dissolve at rate $W$, while the remaining are eroded away at rate $E$ without being chemically altered (Riebe et al 2004, Dixon et al 2009). The weathering flux, and hence the partitioning of the denudation flux, is determined by all the factors that control the weathering reaction (Lebedeva and Brantley 2018). These include geochemical (e.g., intrinsic weathering rate), geologic (e.g., denudation rate), as well as hydrologic parameters (e.g., precipitation and potential evapotranspiration (Budyko 1974, Sposito 2017, Daly et al 2019)).

The $\pi$ theorem of dimensional analysis (Barenblatt and Isaakovich 1996) allows us to establish a relation between the dimensionless ratio of weathering to denudation rate $W/D$ (also called chemical depletion fraction) and an array of dimensionless numbers, formed by grouping the parameters affecting the weathering rate according to their dimensions. Accordingly, as detailed in the appendix A, one can write the following general relationship

$$W/D = f(W_{\text{max}}/D, \text{PET}/P, ..),$$

where $W_{\text{max}}/D$ and $\text{PET}/P$ are the two major groups governing $W/D$ globally (see section 3 and appendix), $W_{\text{max}}$ and $\text{PET}$ being the intrinsic reaction rate of the mineral (e.g., moles/year) and the potential evapotranspiration (e.g., mm/year), respectively. The parameter $\text{PET}/P$, commonly called dryness index $D_1$ (or aridity index), is a popular climatic index used to define the partitioning of precipitation into evapo-transpiration and runoff (Budyko 1974, Daly et al 2019) and a good indicator to which to relate plant productivity and microbial activity in response to water availability (Rodriguez-Iturbe and Porporato 2004). $D_1$ thus accounts not only for water availability but also for the supply of CO2 produced by soil respiration (i.e., by vegetation and microbes). The chemical depletion fraction can also be expressed in a form independent of the type of mineral, $(W/D)$*, by dividing $(W/D)$ by the maximum value reached for $D_1 \to 0$ (appendix B). In this way, $(W/D)$* reduces to a function of $D_1$ only, which can be studied through an analysis of available data.
3. Wetness as the dominant driver of chemical depletion

Guided by the formulation above, we analyzed measurements of elemental chemical depletion fraction (i.e., \(W/D\)) from four data collections (Chadwick et al 2003, Riebe et al 2004, Rasmussen et al 2011, Dere et al 2013) in relation to the \(D_f\) for each site. We considered only elemental, and not total, chemical depletion fractions because depletion of an individual element can be representative of those minerals bearing that element (White and Brantley 2003) (e.g., Na for plagioclase), hence decreasing the uncertainty due to mineralogy (see figure C1 available online at stacks.iop.org/ERC/2/085005/mmedia). Data include granitic terrains (Riebe et al 2004, Rasmussen et al 2011), basaltic terrains (Chadwick et al 2003) as well as shale (Dere et al 2013), and encompass a broad range of environmental settings.

For each experimental site of the data collection, we calculated the dryness index from long-term averages of precipitation and potential evapotranspiration. Measurements of long-term precipitation were provided by all settings with similar dimensionless groups becomes important when focusing on individual locations or when comparing different sites with similar \(D_f\) (figure 1).

\[ (W/D)^* = 1 - \frac{\ln(D_f^a + 1)}{1 + \ln(D_f^a + 1)} \]  

where the only empirical constant is the exponent \(a\). As can be seen in figure 1, the data are in good agreement with the expression (2) above; the root-mean-square error (RMSE) is \(0.19\) and \(\approx 70\%\) of the variance is explained (figure C2). The characteristic exponent of the dryness index is \(a \approx 2.5\). This good agreement underlines the role of wetness in controlling weathering rates at the global scale, while the dependence on other dimensionless groups becomes important when focusing on individual locations or when comparing different sites with similar \(D_f\) (figure 1).
Equation (2) also provides useful insight into how wetness affects the chemical depletion fraction. In particular, it emphasizes the strongly nonlinear relation between water availability and weathering rate. Starting from high $D_I$ values, $(W/D)^*$ increases only slightly up to $D_I \approx 2$, after which a steep increase begins before plateauing at $D_I < 0.5$. The dryness index of approximately 2, from which the strongly nonlinear enhancement of weathering initiates, corresponds to the establishment of grasslands, savannas, and shrublands ecosystems (Woodward et al 2004), emphasizing the important role of vegetation in acidifying the soil and in turn enhancing weathering rates. Interestingly, a similar nonlinear dependence on wetness was observed in the global distribution of organic carbon sorbed to reactive soil minerals (Kramer and Chadwick 2018).

4. Global patterns of weathering

The relationship between chemical depletion fraction and dryness index can be used to estimate the spatial distribution of chemical depletion fraction over the globe from long-term measurements of potential evapotranspiration and precipitation. We considered the CRU dataset with over a century of climatological data (1901-2018) at 0.5° by 0.5° spatial resolution, from which we retrieved potential evapotranspiration and precipitation, and computed a time average for each grid point (see appendix C). From the global pattern of $D_I$ (figure 2(a)), as expected the tropical regions, e.g., the Amazon, Central Africa, and Indonesia, have the lowest dryness index ($D_I$ approaching 0), mostly due to their high precipitation regime. High latitude regions also have very low dryness index values because of their low potential evapotranspiration. Both subtropical and temperate regions exhibit high levels of dryness, including Northern Africa and Western and Central Asia, as well as west USA, Chile and Southern Argentina, and most of Australia. The relation between $D_I$ and latitude is illustrated in the longitudinal average $D_I$, clearly showing that the subtropics are the driest latitudes.

The resulting distribution of chemical depletion fraction follows an opposite pattern compared to $D_I$, with $(W/D)^*$ values that increase with wetness (i.e., lower $D_I$), but due to the strongly nonlinear relation between $(W/D)^*$ and $D_I$, equation (2), the distinction between regions of slightly and highly depleted soils is more pronounced. Many regions appear to have low values of $(W/D)^*$, while very high values are estimated only in
regions with $D_I \to 0$, such as tropical and northern latitude regions. This is reflected in the longitudinal average (figure 2), which shows the two main peaks at the equator and northern latitudes. The similarity with the global soil pH distribution (IGBP-DIS 1998) further highlights the important role of soil CO$_2$ production and its transport to reaction sites (Calabrese et al 2017) on the weathering regime. In the USA, according to wetness, $(W/D)^s$ increases from west to east coast, except for the Pacific Northwest rainforests, following the same geographical pattern recently observed by Slessarev et al (2019). Our analysis also included Greenland, where despite low temperatures high rates of chemical weathering have been observed (Graly et al 2014, Scribner et al 2015). Here, however, results need to be interpreted cautiously.

Globally, while values of $D_I$ about 0.8 are the most frequent and 80% of the land has $D_I < 3$ (figure 3(a)), the frequency distribution of $(W/D)^s$ has a peak at about 0.15 and one at intermediate values (0.4 to 0.7) (figure 3(c)). This particular frequency distribution of $(W/D)^s$ originates, by conservation of probability, from the frequency distribution of $D_I$, according to which low values are more frequent, and from the nonlinear relation with $(W/D)^s$ and the power law decay with $D_I$. Overall, we found that the global average is $(W/D)^s = 0.5$ and that 50% of the land has $(W/D)^s < 0.5$.

Having an explicit relation between chemical depletion and climate, equation (2), it was possible to define uniquely the weathering regimes by identifying the thresholds between them, for instance by using the points of the curve with maximum curvature, which are located at $D_I$ approximately 0.23 and 0.96. For $D_I < 0.23$, conditions are favorable to the weathering reaction, so that $(W/D)^s \approx 1$ and limitations may arise from the supply of fresh minerals. On the contrary, for $D_I > 0.96$, $(W/D)^s < 0.6$ because conditions are not favorable to the kinetics of the weathering reaction. Weathering regimes could also be defined based on a fixed absolute value of the slope of (2), and essentially the same results would be obtained by fixing a value of the slope of 0.4.

5. Susceptibility to wetness

To identify those areas that potentially can be more impacted by shifts in wetness, we defined the susceptibility to wetness as the negative derivative of equation (2),

$$
\chi = -\frac{d((W/D)^s)}{dD_I} = \frac{\alpha D_I^{\alpha-1}}{(1 + D_I^{\alpha})(1 + \ln(D_I^{\alpha} + 1))^3},
$$

which represents a measure of the sensitivity of different climatic regions and their weathering rate to variability in wetness. It is important to bear in mind that equation (3) does not provide information about timescales, for example of a transition towards a different value of $(W/D)^s$ upon a change in $D_I$. A change in wetness is likely to impact directly the weathering rate, but the extent to which the latter will have an effect on the chemical depletion fraction of a soil profile will also depend on geologic (e.g., mineralogy, denudation, erosion) and surface and subsurface topographic factors, as well as temperature (e.g., (Rempe and Dietrich 2014, Riebe et al 2017, Holbrook et al 2019, Richter et al 2020)). Therefore, changes in long term chemical depletion of soil profiles should be analyzed also with respect to these other factors.

![Figure 3. Statistics of global $D_I$ and $(W/D)^s$. (a) Frequency histogram of global $(W/D)^s$. (b) Functional relationship, equation (2), between $(W/D)^s$ and $D_I$. (c) Frequency histogram of global $D_I$.](image-url)
As can be seen in figure 4, the susceptibility tends to zero for low and high values of $D_D$, while it peaks at intermediate values of $D_D$ ($0.5 < D_D < 1$), showing that the transitional regime also corresponds to the highly susceptible one. For regions with low or high $D_D$, corresponding to supply and kinetic limited areas, respectively, the susceptibility to changes in wetness is very low. In humid regions (low $D_D$) there is more water and acidity than needed to reach the maximum weathering rates, while in dry regions such as desert and semi-desert regions (high $D_D$) water availability does not impact considerably the production of soil CO2 given the poor presence of vegetation and low soil organic matter. Areas in the transitional regime, with intermediate $D_D$, are highly susceptible and could potentially enter the supply- or kinetic-limited regimes depending on whether $D_D$ decreased or increased, respectively.

According to the thresholds in dryness index, 0.23 and 0.96, we divided the land into regions of supply-limited, kinetic-limited or transitional regime (figure 5). Globally, about 61% of land surface is in kinetic limited conditions, including a large portion of north America and Asia, only 1% is in supply limited conditions, and the remaining 38% is in the transitional regime and thus susceptible to changes in wetness (see inset in figure 5). Susceptible regions include the African rainforest, eastern north America, central and northern Europe, as well as continental south-east Asia.

6. Conclusions

In conclusion, the relation between weathering, denudation, and wetness (equation (2)) introduced in this analysis sheds light on the mechanisms through which climate, specifically water availability, governs chemical
that the long-term evapotranspiration can be expressed as a function of mostly driven by climatic factors under consideration here, a common assumption to solve the water balance is that the hydrologic partitioning is the CO2 enrichment of water as it references therein water and minerals as well as the transport of the chemical species reaction rate constant, temperature, and the hydrologic regime, which determines the contact time between denudation rate, which can be considered the input of fresh minerals to the weathering zone, the mineral precipitation, ET

The process of chemical weathering along a soil profile can be described by the balance equations for mineral (Riebe et al 2004) and water (Budyko 1974, Sposito 2017, Daly et al 2019),

\[
D = E + W \\
P = ET + R,
\]

where \(D\) is the denudation rate, \(E\) is the erosion rate, \(W\) is the weathering rate, \(P\) is the long-term average precipitation, \(ET\) is the long-term evapotranspiration, and \(R\) is the long-term runoff. For the long timescales under consideration here, a common assumption to solve the water balance is that the hydrologic partitioning is mostly driven by climatic factors (Budyko 1974), such as precipitation and potential evapotranspiration, \(PET\), so that the long term evapotranspiration can be expressed as a function of \(P\) and \(PET\), i.e.,

\[
ET = ET(P, PET),
\]

\(PET\) being the long-term average potential evapotranspiration. The weathering flux is a function of the denudation rate, which can be considered the input of fresh minerals to the weathering zone, the mineral reaction rate constant, temperature, and the hydrologic regime, which determines the contact time between water and minerals as well as the transport of the chemical species (e.g., (Dupré et al 2003, Dixon et al 2009) and references therein). The hydrologic regime is also an important driver of vegetation dynamics and microbial activity in the upper soil layers (Rodríguez-Iturbe and Porporato 2004), and hence controls soil respiration and the CO2 enrichment of water as it flows through the soil column towards the weathering zone. To couple the weathering rate to the water balance, we can express the weathering flux as

\[
W = W(D, W_{\text{max}}, P, PET, \ldots),
\]

where \(W_{\text{max}}\) is the maximum weathering rate for the type of mineral. Equation (A.3) is also a function of other factors, such as temperature, activation energy and so on, but as it will be shown later, at the global scale these exert only a second order control on \(W\), so we will not treat these factors explicitly. We can select two dimensionally independent variables in equation (A.3) and reduce it to a relation between fewer dimensionless variables. Selecting the denudation rate, \(D\), and precipitation, \(P\), equation (A.3) can be rewritten as

\[
\frac{W}{D} = f \left( \frac{W_{\text{max}}}{D}, \frac{PET}{P}, \ldots \right),
\]

where \(f\) is the scaling relationship between the ratio of weathering and denudation flux, \(W/D\), referred to as the chemical depletion fraction (Riebe et al 2004), the ‘potential’ chemical depletion fraction, \(W_{\text{max}}/D\), corresponding to the chemical depletion fraction obtained if the reaction proceeded at the maximum rate, and the ratio of potential evapotranspiration and precipitation, \(PET/P\), referred to as the dryness index \(D_t\) (Budyko 1974).

As for the expression of a chemical reaction rate, consisting of the product of the intrinsic rate constant and another term which is function of the saturation, here we expect the maximum reaction rate to be a multiplicative factor for the weathering rate. Accordingly, in this dimensionless framework equation (A.4) can be written as
\[
\frac{W}{D} = f_1\left(\frac{W_{max}}{D}\right) \cdot f_2\left(\frac{\text{PET}}{P}, \ldots\right),
\]

where, because \(W/D\) varies between 0 and 1, both functions \(f_1\) and \(f_2\) also vary between 0 and 1. The chemical depletion fraction \(W/D\) is equal to \(f_1\) when \(D_1 \to 0\), according to which \(f_1 \to 1\). For this limit, \(W/D\) in fact reduces to a function only of \(W_{\text{max}}/D\) (figure C1). For \(W_{\text{max}}/D < 1\), denudation rates are high compared to the maximum weathering rate \(W_{\text{max}}\), so we have \(W = W_{\text{max}}\) and \(W/D = W_{\text{max}}/D\). For denudation rate lower than the maximum weathering rate, \(W_{\text{max}}/D > 1\), the input of minerals from regolith production could not fully support weathering, so that \(W = D\) and hence \(W/D = 1\).

Appendix B. Definition of specific chemical depletion fraction

To focus on the relationship between chemical weathering and climate, it is convenient to focus on the function \(f_2\). This is done simply by dividing \(W/D\) by \(f_1\), so as to obtain

\[
(W/D)^* = \frac{(W/D)}{f_1} = f_2(D_1).
\]

To study the relation between chemical depletion and \(D_1\) from data, the chemical depletion measurements for given element were divided by the value of the maximum chemical depletion fraction, corresponding to the average value of \(W/D\) for the sites whose \(D_1 \to 0\) (see figure C1). Specifically, for each element we considered sites with \(D_1 < 0.5\) and averaged the corresponding \(W/D\) values (for Na, we removed one outlier; \(W/D = 0.25\) at \(D_1 = 0.18\)). Consistently with observed mobility (White and Buss 2014), we obtained \(f_1^{\text{Ca}} \approx f_1^{\text{Na}} > f_1^{\text{Si}}\). Furthermore, \(f_1\) was greater for basalt, reflecting the higher weatherability of basaltic lithologies (Dupré et al 2003). By dividing the \(W/D\) values by the corresponding \(f_1\), we obtained the specific chemical depletion fraction, which does not depend on the type of mineral. The explicit functional form of \(f_2\) is shown in equation (2) in the main text.

As we mentioned above, the analysis explicitly accounts for \(W_{\text{max}}/D\) and \(D_1\), as these are primary factors affecting \(W/D\) at the global scale. Arhenius plots for \(D_1\)-corrected \((W/D)^*\) indeed do not show any strong relation with \(T\) (figure C3 available online at stacks.iop.org/ERC/2/085005/mmedia). Temperature however still indirectly affects water availability through \(\text{PET}\). While, we do not have enough information on the field sites to explore the dependence of \((W/D)^*\) on all possible factors, our hypothesis of the primary control of \(D_1\) is confirmed by its large explanatory power, i.e., \(D_1\) explains 70% of the variance of \((W/D)^*\).

Appendix C. Global distribution of \(D_1\) and \((W/D)^*\)

To compute the dryness index for the globe, we retrieved precipitation and potential evapotranspiration data from the CRU dataset (Harris et al 2014), containing over a century of climatological data (1901–2018) at a spatial resolution of 0.5° by 0.5°. For each grid point, we computed the temporal average for both precipitation and potential evapotranspiration over the period 1901–2018. The dryness index was then computed for each grid point by dividing the average potential evapotranspiration by the corresponding average precipitation. For given latitude, the longitudinal average of \(D_1\) was calculated as

\[
\langle D_1 \rangle = \frac{1}{N} \sum_{i=1}^{N} D_1(i),
\]

where \(\langle D_1 \rangle\) is the longitudinal average of \(D_1\). \(D_1(i)\) is the dryness index at the longitudinal grid cell \(i\) and \(N\) (\(N = 720\)) is the number of cells that cover a circle of constant latitude. This includes the grid cells in the ocean, which however were excluded from the analysis.

For each grid point, \((W/D)^*\) was calculated by applying equation (2) of the main text. For given latitude, the longitudinal average of \((W/D)^*\) was calculated as

\[
\left\langle \left(\frac{W}{D}\right)^* \right\rangle = \sum_{i=1}^{N} \left(\frac{W}{D}\right)^*(i),
\]

where \(\left\langle \left(\frac{W}{D}\right)^* \right\rangle\) is the longitudinal average of \((W/D)^*\), \((W/D)^*(i)\) is the dryness index at the longitudinal grid cell \(i\) and \(N\) is the number of cells that cover a circle of constant latitude (\(N = 720\)).

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Lebedeva M and Brantley S 2018 A clarification and extension of our model of regolith formation on hillslopes Earth Surf. Processes Landforms 43 2715–23
Li D, Jacobson AD and McInerney DJ 2014 A reactive-transport model for examining tectonic and climatic controls on chemical weathering and atmospheric CO2 consumption in granitic regolith Chem. Geol. 365 30–42
Maher K 2010 The dependence of chemical weathering rates on fluid residence time Earth Planet. Sci. Lett. 294 101–10
Maher K and Chamberlain C 2014 Hydrologic regulation of chemical weathering and the geologic carbon cycle Science 343 1502–4
Maher K and Navarre-Stitcher A 2019 Reactive transport processes that drive chemical weathering: from making space for water to dismantling continents Rev. Mineral. Geochem. 85 349–80
Minsny B and Mcbratney A R 2001 A rudimentary mechanistic model for soil formation and landscape development: li. a two-dimensional model incorporating chemical weathering Geoderma 103 161–79
Moon S, Perron J, Martel S, Holbrook W and St Clair J 2017 A model of three-dimensional topographic stresses with implications for bedrock fractures, surface processes, and landscape evolution Journal of Geophysical Research: Earth Surface 122 823–46
Oliva P, Viere J and Dupré B 2003 Chemical weathering in granitic environments Chem. Geol. 202 225–56
Pelak N F, Parolari A J and Porporato A 2016 Bistable plant-soil dynamics and biogenic controls on the soil production function Earth Surf. Processes Landforms 41 1011–7
Rasmussen C et al 2011 Strong climate and tectonic control on plagioclase weathering in granitic terrain Earth Planet. Sci. Lett. 301 521–30
Rempe D M and Dietrich W E 2014 7.4-natural weathering rates of silicate minerals White A and Buss H 2014
Saltzman B 2001 Dynamical Paleoclimatology: Generalized Theory of Global Climate Change Cambridge, UK: Cambridge University Press p 272 June 2001
Riebe C S, Hahn W J and Brantley S L 2017 Controls on deep critical zone architecture: a historical review and four testable hypotheses Earth Surf. Processes Landforms 42 128–56
Riebe C S, Kirchner J W and Finkel R C 2003 Long-term rates of chemical weathering and physical erosion from cosmogenic nuclides and geochemical mass balance Geochim. Cosmochim. Acta 67 4411–27
Riebe C S, Kirchner J W and Finkel R C 2004 Erosional and climatic effects on long-term chemical weathering rates in granitic landscapes spanning diverse climate regimes Earth Planet. Sci. Lett. 224 547–62
Rodriguez-Iiturbe I and Porporato A 2004 Ecohydrology of Water-Controlled Ecosystems: Soil Moisture and Plant Dynamics (Cambridge: Cambridge University Press)
Saaflink M W, Ayora C and Carrera J 1998 A mathematical formulation for reactive transport that eliminates mineral concentrations Water Resour. Res. 34 1649–56
Saltzman B 2001 Dynamical Paleoclimatology: Generalized Theory of Global Climate Change (Amsterdam: Elsevier)
Scribner C A, Martin E, Martin J, Deuerling K, Collazo D and Marshall A 2015 Exposure age and climate controls on weathering in deglaciated watersheds of western greenland Geochim. Cosmochim. Acta 170 157–72
Singh J S and Gupta S 1977 Plant decomposition and soil respiration in terrestrial ecosystems The Botanical Review 43 449–528
Slessarev E, Feng X, Bingham N and Chadwick O 2019 Landscape age as a major control on the geography of soil weathering Global Biogeochem. Cycles 33 1513–31
Sposito G 2017 Understanding the budyko equation Water 9 236
West A J, Galy A and Bickle M 2005 Tectonic and climatic controls on silicate weathering Earth Planet. Sci. Lett. 238 211–28
White A F and Blum A E 1995 Effects of climate on chemical weathering in watersheds Geochim. Cosmochim. Acta 59 1729–47
White A F and Brantley S L 2003 The effect of time on the weathering of silicate minerals: why do weathering rates differ in the laboratory and field? Chem. Geol. 202 479–506
Winnick M J and Maher K 2018 Relationships between CO2, thermodynamic limits on silicate weathering, and the strength of the silicate weathering feedback Earth Planet. Sci. Lett. 485 111–20
Woodward F I, Lomas M R and Kelly C K 2004 Global climate and the distribution of plant biomes Philosophical Transactions of the Royal Society of London. Series B: Biological Sciences 359 1465–76
Yeh G-T and Tripathi V S 1991 A model for simulating transport of reactive multispecies components: model development and demonstration Water Resour. Res. 27 3075–94