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NITROGEN CYCLE

Convergent evidence for widespread rock nitrogen sources in Earth's surface environment

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Nitrogen availability is a pivotal control on terrestrial carbon sequestration and global climate change. Historical and contemporary views assume that nitrogen enters Earth's land-surface ecosystems from the atmosphere. Here we demonstrate that bedrock is a nitrogen source that rivals atmospheric nitrogen inputs across major sectors of the global terrestrial environment. Evidence drawn from the planet's nitrogen balance, geochemical proxies, and our spatial weathering model reveal that ~19 to 31 teragrams of nitrogen are mobilized from near-surface rocks annually. About 11 to 18 teragrams of this nitrogen are chemically weathered in situ, thereby increasing the unmanaged (preindustrial) terrestrial nitrogen balance from 8 to 26%. These findings provide a global perspective to reconcile Earth's nitrogen budget, with implications for nutrient-driven controls over the terrestrial carbon sink.

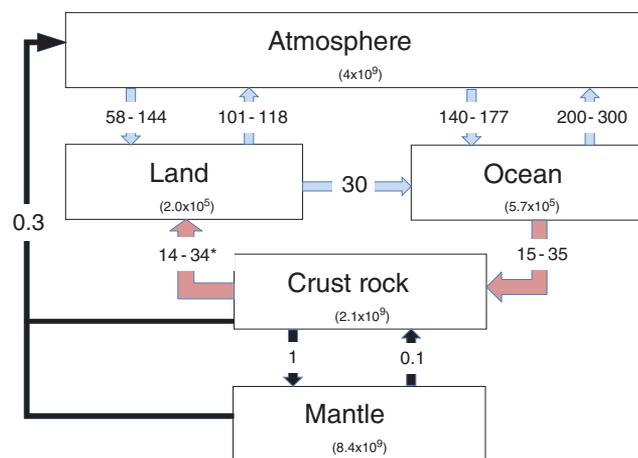
Nitrogen (N) availability controls many aspects of ecosystem structure and function on land and in the sea (1, 2). This includes strong, N-driven effects on Earth's climate system and the size and sustainability of the terrestrial carbon (C) sink (3). Disagreements exist, however, over the biosphere's N balance and how natural sources of N could alter terrestrial C uptake patterns in the future (4). Textbook paradigms and global computational models assume that ecosystems rely principally on the atmosphere for N (5), yet N accumulation rates in vegetation and soil can greatly exceed inputs from biological N fixation and N deposition (6). Recent evidence has raised questions about the role of rock N sources in resolving this discrepancy (7), with potentially widespread implications, given the massive amount of fixed N in the global rock reservoir (8, 9).

We investigated rock N weathering rates in Earth's surface environment, where terrestrial plants, soils, and microbes interact. Over billions of years of Earth history, N has accumulated in rocks, largely as a product of N fixation by aquatic and terrestrial organisms that becomes trapped in sedimentary basins; this N has been traced back to ancient biogeochemical processes, as opposed to contemporary N fixation by free-living microbes and root-associated symbionts. The amount of N varies widely among general rock types; sedimentary and metasedimentary lithologies occupying ~75% of Earth's surface have concentrations of ~500 to 600 mg N kg⁻¹ rock, whereas more spatially restricted igneous rocks often have much lower values (<100 mg N kg⁻¹ rock) (9–11). Although N-rich sediments are globally widespread, such rock N concentrations do

Fig. 1. The preindustrial planetary nitrogen cycle. Fluxes (arrows) in teragrams per year and reservoir sizes in teragrams.

Maintenance of the atmospheric N reservoir requires a transfer of N from crustal rocks to land because degassing fluxes from the crust and mantle are unable to balance N transfer from the ocean to the crust. *Mass required to balance the marine N burial term minus mantle and volcanic degassing. Nitrogen

fixation estimates are from (4, 13, 36, 47); preindustrial nitrogen deposition is from (37); denitrification, hydrologic N transport, and marine burial are from (13, 16); N transport to the mantle and volcanic N emissions are from (14); and N mass reservoirs are from (8, 9). Section 1 of the supplementary materials provides further details.



not translate directly to N inputs in Earth's surface environment. Rather, rock N availability to terrestrial soils and vegetation is determined by denudation (physical plus chemical weathering), which varies as a general function of geochemistry, relief, tectonic uplift, climate, and biology (12).

Therefore, we examined the mobility and reactivity of rock N sources in Earth's surface environment by means of three diverse and largely independent assessments: the planetary mass balance of N (case 1); global-scale denudation and chemical weathering proxy data (case 2); and a spatially explicit N weathering model that uses a statistical probability approach (case 3) (table S1). When combined, these approaches enable us to identify a hitherto unrecognized source of rock N that is ecologically important across Earth's diverse environments and at the planetary scale.

Rock N weathering and the missing N in the planetary balance (case 1)

A classical approach for gaining insight into the magnitude of Earth's biogeochemical transfers relies on the principle of mass and energy conservation. By accounting for N inputs and outputs among Earth system reservoirs, biogeochemists can draw coarse-scale inferences about N exchanges between the atmosphere, biosphere, hydrosphere, and geosphere, which can then be vetted against evidence from more direct approaches. Whereas traditional “box and arrow” models have emphasized N transfers between land, air, and water systems (13), recent quantitative advances in Earth system modeling point to sustained transport of sedimentary marine N through the deep Earth (i.e., the mantle) (14). Considering the geochemical and biological fluxes together illuminates a discrepancy in the planetary N balance, which can be resolved by considering the return of rock N to the land-surface environment (Fig. 1).

Models of the planet's N balance can be classed into two domains: (i) short-term models that emphasize N transformations and fluxes between the atmosphere, biosphere, and human activities (13) and (ii) longer-term models that consider N exchange between the atmosphere and mantle (15). Short-term perspectives emphasize how atmospheric N inputs are balanced by physical and microbial denitrification processes, which return N back to the atmosphere, and thereby ignore the return of N from rocks to land. Instead, these models treat marine sedimentary burial as a global N sink with an integrated flux of ~15 to 35 Tg N year⁻¹ (13, 16).

However, over the longer term, the amount of N entering marine sediments should reasonably balance the amount of N leaving this reservoir. Tectonic uplift forces sediments to Earth's surface over millions of years, a mass-transfer flux that is ultimately balanced by denudation. A fraction (<10%) of sedimentary N enters the mantle, where it can be stored or returned to the atmosphere via

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volcanic emissions [combined long-term average, $\sim 1.5 \text{ Tg N year}^{-1}$ (14)]. Hence, the evidence suggests that considerably more N is entering the rock N reservoir than can be accounted for by mantle exchange and volcanic emissions—what can explain this discrepancy?

One possibility involves non-steady-state conditions, wherein N accumulates in crustal rock and the mantle indefinitely. A simple calculation shows that this explanation is improbable: The atmospheric N_2 reservoir would have completely vanished after <300 million years in this case (17). Although variations in N transfers over geological epochs (e.g., the Deccan Traps) could alter this calculation, for much of the Phanerozoic eon, marine organic matter burial (and hence organic N burial) is estimated to have fluctuated by less than a factor of 2 (18); thus, long-term volcanic degassing and mantle advection fluxes would need to increase by a factor of 20 to 40 over modern rates to compensate for sedimentary N burial. Such increases are unlikely given that sustained volcanism of this magnitude would lead to mass extinctions on land and severe ocean acidification (19).

Instead, the most parsimonious explanation centers on mass-balance closure in which the substantial transfer of N from rocks to the land surface is balanced by N burial rates in the seafloor (Fig. 1). This alternative fits with evidence for a stable atmospheric N_2 reservoir over the Phanerozoic (2) and is aligned with Earth's dynamic rock cycle, whereby sediments are lithified and tectonically uplifted to replenish losses from continental erosion (Fig. 1). Although humans have both purposely and inadvertently increased the terrestrial biosphere's N balance (13), this modern-day perturbation is insubstantial vis-à-vis the cumulative N cycle transfers that have taken place over Earth history. Thus, the planetary-scale mass balance points to a nontrivial N weathering term from the continental rock reservoir of ~ 14 to $34 \text{ Tg N year}^{-1}$ (equation S2).

Evidence for rock N denudation and weathering based on global proxies (case 2)

The planet's N inventory provides coarse-scale evidence for substantial N weathering reactions in Earth's surface environment, but this approach is more impressionistic than direct. Likewise, our extensive sampling efforts (11) and other global-scale syntheses (9, 10) reveal widespread rock N sources in the near-surface environment (Fig. 2A); however, these results do not address global N weathering rates. We must place quantitative constraints on rock N denudation rates to draw more direct inferences about N chemical weathering inputs worldwide. This includes an analysis of the organic N that is bound in sedimentary rocks and the mineral N in silicates (largely as NH_4^+).

We first consider global-scale constraints on organic N denudation rates (physical plus chemical weathering) in sedimentary rock by combining data on organic C denudation rates with the C:N stoichiometry of sedimentary rock (11). Fossilized organic C denudation rates range between 100 and $143 \text{ Tg C year}^{-1}$ (20) and show

good agreement with estimates of global C burial in marine sediments [126 to $170 \text{ Tg C year}^{-1}$ (21–23)]. Dividing these fluxes by the average C:N ratio of sedimentary rock (8.13 by mass; supplementary materials) yields a N denudation flux between ~ 12 and $18 \text{ Tg N year}^{-1}$ and a marine burial rate of ~ 16 to $21 \text{ Tg N year}^{-1}$ (Table 1). These global N denudation values agree reasonably well with the planetary mass balance (discussed above), despite the very different data and techniques used in each case (table S1).

A second approach derives from basin-scale sediment and solute fluxes and cosmogenic radionuclide (CRN) denudation data, which collectively reflect the net movement of silicate rocks from the terrestrial to marine environment. This

approach addresses the N in mineral form. Using sediment and solute budgets, Milliman and Farnsworth (24) estimate that $\sim 23 \text{ Pg}$ of total silicate-rock mass is delivered to the global oceans annually. Combined with endorheic basins (environments not in contact with the ocean), the global land-to-sink mass flux of silicates is roughly 28 Pg year^{-1} . Applying our global mean N lithology concentration of 337 mg N kg^{-1} to the mass flux of silicates (supplementary materials) yields a N denudation flux of $\sim 9 \text{ Tg N year}^{-1}$ (Table 1) for silicate-bound N.

The results of this calculation are consistent with findings from catchment-scale CRN analysis, which suggest a global rock denudation flux of 28 Pg year^{-1} (25); however, grid-scale biases

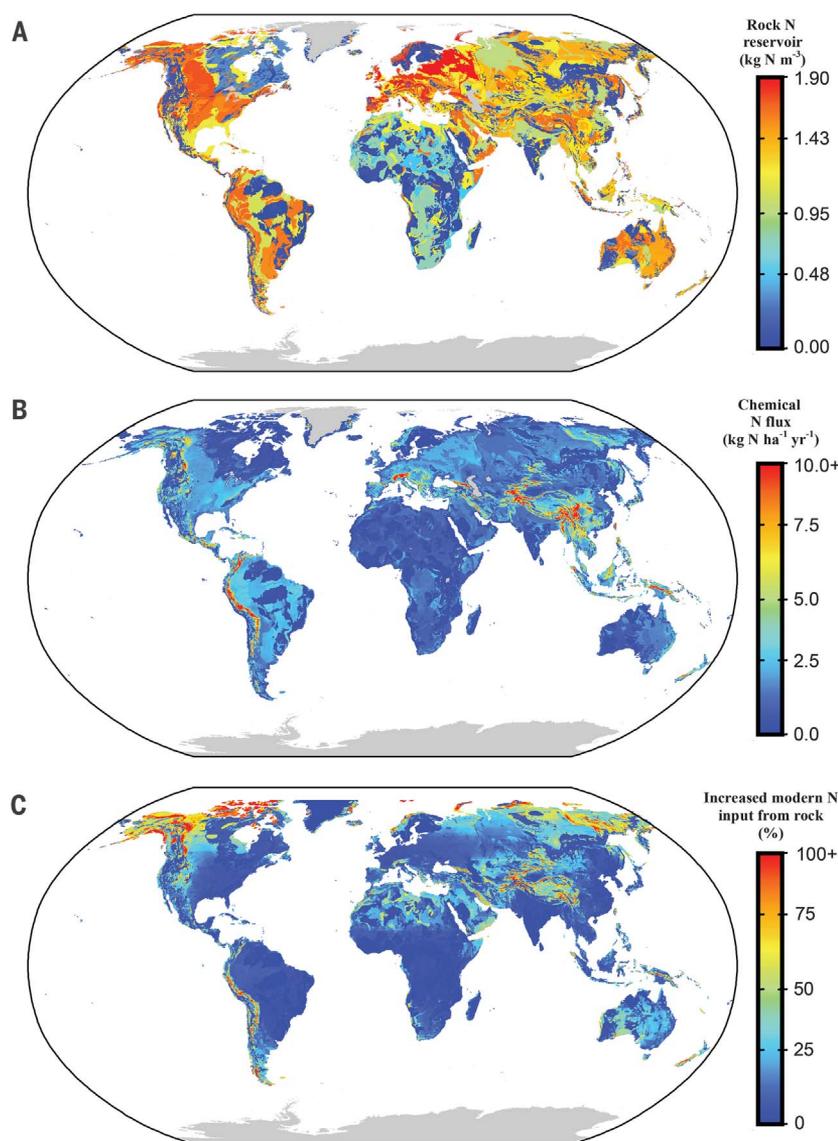


Fig. 2. Spatial patterns of rock N concentrations, weathering fluxes, and rock N contributions versus atmospheric sources of N. (A) Spatially explicit estimates of surface (top 1 m) rock N concentrations. (B) Nitrogen chemical weathering fluxes derived from our globally calibrated model. (C) Percent increase in the preindustrial terrestrial N balance with rock N inputs (i.e., rock N weathering divided by the sum of atmospheric N inputs). The model points to highest absolute (B) and relative (C) rock N input fluxes in mountainous regions and at higher latitudes.

Table 1. Estimates of global rock nitrogen reservoirs and fluxes.

	Case 2			Case 3
	Fossilized organic matter	Silicate solute and sediment yield	Catchment-scale CRN	Global spatial model
N reservoir (petagrams N in top 1 m of rock)	149	N/A	N/A	92–110
N denudation (teragrams N per year)	12–18	9	2–22	19–31
Marine burial (teragrams N per year)	16–21	N/A	N/A	N/A
Terrestrial chemical weathering (teragrams N per year)	5–12*	0.9–4.5	0.7–11†	11–18‡

*Range of organic nitrogen weathering. †Range of mineral nitrogen weathering. ‡Organic plus mineral nitrogen weathering from the spatial model.

and extrapolation (26) of catchment-scale studies to Earth's surface produce a broad range of estimates at the global scale (6 to 64 Pg year⁻¹). On the basis of these end-members and the mean global N reservoir above, the silicate-rock N denudation flux may be as low as 2 and as high as 22 Tg N year⁻¹. Importantly, such CRN-derived estimates do not consider the acceleration of erosion and denudation rates through modern land-use practices, which have increased erosion by a factor of 10 to 100 (27).

The chemical weathering quotient of the total N denudation fluxes can be estimated from fossilized organic matter (FOM) weathering and chemical depletion of silicates. Chemical weathering of FOM occurs more completely than for silicate minerals, because the former is susceptible to oxidation as opposed to kinetically constrained acid-hydrolysis reactions. Globally, chemical (oxidative) weathering of C in FOM varies between 40 and 100 Tg C year⁻¹ (20), which translates to an organic N weathering flux of 5 to 12 Tg year⁻¹ (Table 1 and supplementary materials). In contrast, chemical weathering of silicate rocks is less certain; it varies across parent material, climate, relief, and biological communities. Chemical depletion of silicates in bulk rocks has been shown to vary from 10 to 16% (24, 28), which would imply a mineral N weathering flux between 0.2 and 3.5 Tg year⁻¹ [i.e., the range of N denudation estimates (2 to 22 Tg N year⁻¹; Table 1) multiplied by 0.1 to 0.16].

However, this lattermost calculation does not consider the differential chemical reactivity of elements in rock, which can be particularly rapid for certain rock-derived elements, including N (7, 29). For example, chemical depletion of parent materials varies from virtually nil (e.g., zirconium and titanium) to relatively rapid (e.g., calcium) in common rock substrates and is biased by the presence of quartz, which is highly resistant to chemical weathering. Application of data from our field studies (29) in rapidly denuding mountains of the northern California Coast Ranges demonstrates a N chemical depletion of ~36 to 50%, which raises the silicate N weathering flux to 0.7 to 11 Tg N year⁻¹ (range of weathering for catchment-scale CRN; Table 1).

These diverse geochemical proxies point to a global N denudation flux (organic plus mineral N) that varies between 14 and 40 Tg year⁻¹, with a chemical N weathering fraction between ~6 and 23 Tg N year⁻¹ (weathering range of FOM plus catchment-scale CRN; Table 1). These estimates confirm expectations from the planetary mass-balance results (case 1) and derive from actual proxies of physical and chemical weathering within a given set of assumptions; however, neither case 1 nor case 2 address spatial patterns and environmental controls on N weathering rates across Earth's surface—issues that we addressed with our spatial global weathering model.

A probabilistic modeling approach to N weathering inputs worldwide (case 3)

We developed a data-driven modeling approach to spatially quantify global N weathering fluxes. Our model incorporates topographic, climatological, and lithological factors to estimate N denudation and chemical weathering rates, and it is calibrated using solute sodium (Na⁺) fluxes from 106 large river basins across Earth (30). It differs from previous approaches in that we rely on machine-learning algorithms, quantile regression, and Monte Carlo simulations, as opposed to the more classical mean-field parameterization schemes. We applied our model at 1-km²-grid scales, using mass-balance equations developed at hillslope to small basin scales (31). The conservation-of-mass equations used in our model take the form

$$D_{N,Na} = (Q_D)(\rho)([N, Na]_{\text{rock}}) \quad (1)$$

$$CDF_{Na} = 1 - \frac{[Na]_{\text{soil}}}{[Na]_{\text{rock}}} \quad (2)$$

$$W_{Na} = (D_{Na})(CDF_{Na}) \quad (3)$$

$$W_N = (D_N)(f_{OM-N})(CDF_{org-N}) + (D_N)(1 - f_{OM-N})(CDF_{Na}) \quad (4)$$

where $D_{N,Na}$ (mass × length⁻² × time⁻¹) is the element-specific (N or Na⁺) denudation flux, Q_D is the denudation rate (length × time⁻¹), ρ is rock density (mass × length⁻³), and $[N, Na]_{\text{rock}}$ is the element-specific concentration in rock

(mass × mass⁻¹). Chemical depletion of Na⁺ from silicate rocks (CDF_{Na}) is applied to both Na⁺ and N weathering functions (section 3 of the supplementary materials). W (mass × length⁻² × time⁻¹) is the element-specific (N or Na⁺) chemical weathering flux, and f_{OM-N} (dimensionless) is the fraction of total rock N in organic forms.

Briefly, our model relies on Monte Carlo methods to estimate probability values for Q_D , $[N, Na]_{\text{rock}}$, and CDF_{Na} , with 10,000 simulations per parameter per cell. We calibrated the model by minimizing residuals between the modeled and empirically observed basin-scale Na⁺ training set (W_{Na}). We estimated denudation (Q_D) by using a statistical model that incorporates catchment-scale CRN denudation rates (32) and digital topography. Rock N and Na⁺ concentrations ($[N]_{\text{rock}}$ and $[Na]_{\text{rock}}$) were derived from our synthesis of measurements (11) and the U.S. Geological Survey geochemical database, respectively (33). We used a generalized additive model to estimate the chemical depletion fraction (CDF_{Na}). The factors in the model include topographic relief, evapotranspiration, and excess water (precipitation minus evapotranspiration) (supplementary materials). We parameterized the CDF model by using 41 separate observations of soil Na⁺ depletion rates collected from the primary literature (section 3 of the supplementary materials).

These simplifying assumptions capture generalized patterns of chemical weathering rates as a function of climate and topographic relief, as calibrated with salt-corrected riverine Na⁺ fluxes to the ocean (tables S5 and S6). The model's reliance on soil-based chemical depletion rates is limited in low-relief landscapes, in areas where subwatershed measurements may be decoupled from larger-scale fluxes, and in recently deglaciated terrains (28). Yet our simulations are consistent with general global observations of soil development and weathering patterns (figs. S3 and S4) and the anticipated switch from supply-limited to transport-limited kinetics in chemical weathering that has been observed for high-relief landscapes (34). Further, total rock denudation rates predicted by our model (46 to 61 Pg year⁻¹) fall within the range of previous studies [20 to 64 Pg year⁻¹ (25)].

At the global scale, our model simulates a large N denudation flux, consistent with cases 1 and 2. Specifically, we estimate that ~19 to 31 Tg N year⁻¹ is denuded from the land-surface environment, with a chemical weathering flux of 11 to 18 Tg N year⁻¹ (Table 1 and Fig. 2B). These results suggest that ~40 to 60% of rock N is chemically released to the terrestrial surface environment before export, consistent with field studies of mineral N depletion rates in mountainous areas (29); that is, ~50% of rock nitrogen is lost to physical erosion without entering terrestrial ecosystem pools in situ. We do not consider the fate of such physically eroded N in downslope ecosystems, which would likely increase the global N weathering flux in low-relief environments.

The scaled-up spatial N chemical weathering flux corresponds well with mean-field geochemical proxies (Table 1). Furthermore, our geospatial model indicates that as much as ~65% (7 to 12 Tg N year⁻¹) of the total rock N chemical flux is derived from organic N, similar to the FOM-based estimates (5 to 12 Tg N year⁻¹; Table 1). These results appear reasonable given our limited understanding of differences in weathering processes among FOM and silicate rocks.

Across the land surface, rock N weathering is relatively widespread, with variations in N geochemistry, relief, and climate determining the magnitude of rock N inputs to terrestrial ecosystems (Fig. 2B). For example, large areas of Africa are devoid of N-rich bedrock and have relatively low relief and arid climate conditions, which together substantially limit N weathering fluxes. In contrast, some of the highest rock N inputs are estimated for the northern latitudes (Fig. 2B), where N-rich rocks and high-relief landscapes are more prevalent. At regional scales, mountainous regions with high uplift and adequate moisture—for example, the Himalaya and

Andes mountains—are estimated to be large sources of N weathering inputs to land-surface environments, similar to the importance of these regions to global weathering rates and climate (35).

Context and implications

The body of evidence points to substantial rock N denudation and weathering rates at regional to global scales. Although each of our approaches is rooted in mass-balance principles, the diversity of techniques confers a reasonable degree of independence among the case studies (table S1); this adds robustness to the working conclusion of widespread rock N inputs in terrestrial surface ecosystems. Our geospatial model provides the most direct and geographically rich set of predictions, with the global range in fluxes largely driven by the calibration approach (basin- versus global-scale; supplementary materials). Results from the other case studies overlap with the spatial model, and we make conservative assumptions about rock N weathering rates in general (table S1). Future work could therefore cause the case studies to diverge, but with a tendency toward higher rather than lower overall rock N fluxes. We conclude that our findings extend previous plot-scale evidence for rock N weathering inputs in select ecosystems to a global biogeochemical paradigm, and that they indicate considerable limitations in contemporary models, which exclude the role of rock N sources in governing global-scale patterns of terrestrial N availability.

To further examine the importance of rock N weathering vis-à-vis the terrestrial N balance, we compare our geospatial model estimates with N fixation and deposition inputs to natural biomes (i.e., nonagrarian areas; Fig. 2C and Table 2). Isotopically constrained global terrestrial N fixation varies from 58 to 100 Tg N year⁻¹ (36), with

N deposition rates in preindustrial and modern nonagrarian environments varying from 11 to between 30 and 34 Tg N year⁻¹, respectively (37, 38). Thus, although anthropogenic activities have dramatically increased global N inputs through deposition, nearly half of this input is in agricultural and urban landscapes where rock is not likely to be a substantial component of ecosystem N cycling (table S6).

Our findings for rock N weathering rates increase the preindustrial terrestrial nitrogen budget by 8 to 26% (Table 2), with a modern-day rock N contribution to natural systems of 6 to 17% of total N inputs. These calculations point to rock inputs increasing the mean (midpoint) global N budget by 17 and 11% for preindustrial and modern periods, respectively, with more pronounced effects at the biome and regional scales.

Our results show that rock N inputs may be particularly important in montane ecosystems where denudation rates are rapid (Fig. 2C) and high-latitude ecosystems where high biological N fixation rates are temperature-limited (39). Spatially, our analysis suggests that rock N inputs can account for a substantial fraction of modern N inputs (including anthropogenic N deposition) to temperate and montane grasslands (8 to 32%), temperate and boreal forests (9 to 38%), tundra (23 to 51%), deserts (11 to 23%), and Mediterranean shrub- and woodlands (9 to 22%) (Table 2 and table S6). In contrast, rock N inputs constitute a substantially smaller fraction of N inputs to tropical grasslands (2 to 8%) and tropical forests (4 to 12%), where weathering is supply-limited and N fixation rates are naturally high.

Where N weathering occurs deep beneath the soil and regolith, some or all of the N may be released to groundwater and transported to fluvial systems (40–42). Under this scenario, the ability of terrestrial plant communities to use

Table 2. External nitrogen inputs from rock and atmospheric sources (teragrams per year). Values in parentheses show the range of estimates, where available. Biome areas are from World Wildlife Fund ecoregions. Biome-specific N fixation estimates are from (4, 40, 47). Atmospheric N deposition is derived for 1860 (preindustrial) and 1993 from (37) and for 2001 from (38). Global-scale estimates (bottom row) include N inputs from sources above and estimates from (36). N inputs to modern agrarian lands are not considered in these calculations; table S7 shows the agrarian influence.

Biome	Rock weathering		Biological fixation		Atmospheric deposition		
					Preindustrial	Modern	
Tundra	0.7	(0.6–0.9)	0.6	(0.7–1.7)	0.1	0.3	(0.3–0.4)
Taiga (boreal forest)	1.3	(1–1.6)	1.2	(1.6–5.3)	0.6	1.9	(1.3–2.5)
Temperate forest	2.8	(2.1–3.4)	6	(8–8.9)	2.6	5.3	(5.2–5.4)
Temperate grassland	1.2	(0.9–1.4)	1.4	(1.8–3.8)	1	1.9	(1.8–2.0)
Montane grassland	1.2	(0.9–1.5)	1.2	(1.5–2)	0.4	1.9	(1.9–1.8)
Mediterranean	0.3	(0.3–0.3)	0.3	(0.5–1.1)	0.2	0.4	(0.4–0.4)
Deserts	2.1	(1.2–2.5)	7.7	(5.5–10.2)	1	4.7	(4.2–5.1)
Mangroves	0	(0–0)	0.3	(0.3–0.4)	0	0.1	(0.1–0.1)
Tropical forest	3.9	(3–4.7)	40.1	(25–53.1)	3.7	8.1	(7.6–8.7)
Tropical grassland	1.4	(1.1–1.7)	39.2	(14.6–51.9)	1.4	7.0	(6.6–7.3)
Tropical wet grassland	0.1	(0.1–0.1)	1.8	(0.7–2.4)	0.1	0.4	(0.4–0.4)
Rock/water	0.1	(0–0.1)	0.3	(0.2–0.4)	0.1	0.2	(0.2–0.2)
Global totals	15.1	(11.2–18.2)	100.1	(58.0–132.5)	11.2	32.2	(29.9–34.4)

deeply weathered N is dependent on plant-root proliferation into the deep subsurface (i.e., the depth of the critical zone). Woody plants can effectively penetrate deep regolith, with roots extending tens of meters below the terrestrial surface, in environments ranging from deserts to rainforests (43). Inferential work has pointed to the high mobility of rock N in ecosystems, which can be depleted from minerals at rates that exceed Na^+ and K^+ release from silicates (29). The role of microbes may be particularly important in this regard; so-called “rock-eating” fungi can accelerate weathering rates of minerals harboring biologically important nutrients, such as phosphorus (P), K^+ , and Ca^{2+} (44, 45).

Lastly, the availability of N singly and in combination with P profoundly limits terrestrial C storage, with nontrivial effects on global climate change (4, 46). Our previous work demonstrated a doubling of ecosystem C storage among temperate conifer forests residing on N-rich bedrock (7). Our model indicates that rock N inputs could make up >29% of total N inputs to boreal forests, which could help to explain the high C uptake capacity observed for this biome and partially mitigate the mismatch of C and N budgets in Earth system models (3). Historically, weathering has been viewed as responsive to CO_2 enrichment and climate change over deep geological time (millions of years) (35). The direct connections that we draw between tectonic uplift, N inputs, and weathering reactions therefore emphasize a role for rock-derived nutrients in affecting the 21st-century C cycle and climate system.

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Convergent evidence for widespread rock nitrogen sources in Earth's surface environment

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Freed from a rocky embrace

Nitrogen availability is a central controller of terrestrial plant growth and, thereby, of the carbon cycle and global climate change. It has been widely assumed that the atmosphere is the main source of terrestrial nitrogen input. Surprisingly, Houlton *et al.* now show that bedrock is just as large a nitrogen source across major sectors of the global terrestrial environment. They used three diverse and largely independent assessments of the nitrogen mobility and reactivity of rocks in the surface environment. These approaches yielded convergent estimates pointing to the equal importance of the atmosphere and bedrock as nitrogen sources.

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