

GEOLOGY

Understanding Soil Time

Susan L. Brantley

Soils constitute the topmost layer of the regolith, the blanket of loose rock material that covers Earth's surface. An open system such as soil or regolith is sustainable, or in steady state, only when components such as rock particles are removed at the same rate they are replenished. However, soils are defined not only by rock particles but also by minerals, nutrients, organic matter, biota, and water. These entities—each characterized by lifetimes in regolith that vary from hundreds of millions of years to minutes—are often studied by scientists from different disciplines. If soils are to be maintained in a sustainable manner (1, 2), scientists must develop models that cross these time scales to predict the effects of human impact.

With respect to the longest time scales, geologists studying Earth's landscapes argue that continents have experienced balanced rates of tectonic uplift and erosion. Thus, the mass of rock particles produced by regolith-forming processes during uplift is balanced by the mass of particles eroded over geological time scales. At steady state, the mass of particles in the regolith “box,” divided by the rate of removal of particles from that box, defines the particle residence time. If regolith is perturbed, the system moves toward a new steady state within a characteristic response time that—for linear systems—equals about 4 times the residence time.

These concepts are exemplified by observations of an undisturbed ridgetop in the Puerto Rican rainforest. At this site, the rate of particle mass loss due to dissolution and erosion (termed total denudation), cast as the rate of lowering of Earth's surface, is 0.04 mm/year (3). This rate is calculated by assuming that the rate of production of cosmogenic nuclides produced by penetration of cosmic rays into the upper 0.6 m of regolith is balanced by loss of these nuclides through denudation. The residence time for particles in this 0.6-m-thick “cosmogenic box” equals 15,000 years [= 0.6 m/(0.04 mm/year)]. If the thickness of this upper soil were perturbed, it would slowly return to its initial state over ~60,000 years.

As measured from cosmogenic isotopes, residence times in the upper 0.6 m of

regolith range from 100 to 100,000 years for soils worldwide, depending on the intensity of tectonic activity (4). However, in the most tectonically quiescent areas of Africa, the cosmogenic technique no longer works, because the residence time of particles may reach hundreds of millions of years.

In contrast to geologists studying landscapes, geochemists interested in the chemical composition of the regolith focus on the response times of minerals. If one could stand on the 10-m-thick regolith at the ridgetop in Puerto Rico for a sufficiently long time, one would observe bedrock fragmenting into particles at 10 m depth that then diminish in size as they move upward and out of the regolith (see the figure). Quartz particles would ascend without disappearing, defining residence times similar to that of the rock particles. In contrast, feldspar dissolves from regolith particles during their trajectory across the lowest 30-cm layer of regolith, defining a residence time of 7500 years (5). More soluble minerals such as calcite can disappear even faster. Residence and response times of minerals, determined on the basis of chemical soil profiles, thus vary from hundreds of millions to hundreds of years.

If, instead of the minerals, the objects of study in the soil are the nutrients fixed from the atmosphere by organisms (6), the time scales of interest are generally shorter. Residence times of 100 to 1000 years are commonly estimated for soil organic matter, but some of this material turns over within 1 to 10 years (7, 8). Residence times can be even shorter for nitrogen (9).

Efforts to maintain soils in a sustainable manner are complicated by interactions among soil components that respond to perturbation at vastly different rates.



Inside the regolith. Weathering continuously replenishes the regolith while erosion removes soil at the surface. Minerals, organic matter, and water move through the regolith on different time scales, complicating efforts to define what sustainable soils are and how they could be maintained.

Scientists studying biota are often interested in the time scales that define how fast one ecosystem succeeds another after a disturbance. Generally, this response time is tens to hundreds of years. In fact, whether an ecosystem can ever reach steady state is a matter of debate. If it is possible, steady state is a complex function of the extent and frequency of disturbances such as fires and insect infestations (10).

The final component of soil considered here, water, responds at the shortest time scales. Water moves both downward (because of meteoric inputs) and upward (because of evapotranspiration mediated by roots that often extend to depths of tens of meters). Water residence times in regolith are measured with stable isotopes to decipher the interplay of “old” and “new” water. These water types are characterized by long or short residence times varying from tens of years to minutes.

When scientists within a discipline study soils, they generally focus on one of these time scales while ignor-

ing faster and slower processes. Learning how soils will change in the future will require observations and models that cross time scales (11). For example, present-day and long-term denudation rates for catchments or soils have been shown to be equal across time scales in some cases, as required for sustainable soils. In other cases, the long-term and present-day denudation rates do not agree, perhaps because of variations in ecosystems, climate, glacial effects, extreme events, or human impact (4, 12).

Another way to bridge time scales is to study chronosequences—soils formed on the same rock type in the same climate but for varying duration of weathering. For slow-weathering, undisturbed chronosequences,

Earth and Environmental Systems Institute, Pennsylvania State University, University Park, PA 16802, USA. E-mail: brantley@eesi.psu.edu

neither ecosystems nor regolith attain steady state; rather, they vary together as a result of the 30 or so bioessential elements mined by biota from rocks. Most important, phosphorus is extracted at depth by organisms, pumped upward, stored in biota and minerals, and recycled. Because phosphorus is lost to groundwater, however, depletion of regolith causes ecosystem degradation over 1000 to 10,000 years (13). Such coupled processes may be manifested in transformation of both above- and below-ground ecosystems as soils cross thresholds related to changes in pH, redox, and nutrient concentration (14, 15). For example, subsurface ecosystems may become increasingly fungi-dominated as soils become phosphorus-limited (13).

The likelihood of crossing important thresholds is high today given the intensity of anthropogenic impact. Human activities have increased the long-term soil erosion rate by about a factor of 30 globally (1). Global agriculture has also caused nutrient depletion, especially in slow-weathering regions such as Africa. Largely to replenish nutrients, humans have doubled the input of

fixed nitrogen into terrestrial ecosystems above prehuman values globally (16). The use of fertilizers replenishes soils but, given the time scale of soil water flow, also causes escape of nutrients and eutrophication in other ecosystems. For example, the transport of dissolved phosphorus from land to oceans has doubled, largely as a result of fertilizer use (17).

The need to maintain soils sustainably is now driving scientists to formulate models that describe not only how soil components react alone, but how they interact with each other in response to tectonic, climate, and anthropogenic forcing within the so-called Critical Zone—the zone extending from the depth of groundwater up to the outer limits of vegetation. Such models will provide the language that can allow scientists to communicate across disciplinary boundaries, but they must be tested across time scales with use of the sediment record, chronosequences, and observations of modern-day fluxes. Just as we use global climate models today to project future climate change, we will eventually be able to use global soil models to project future soil change.

References

1. B. H. Wilkinson, B. J. McElroy, *GSA Bull.* **119**, 140 (2006).
2. P. H. Bellamy *et al.*, *Nature* **437**, 245 (2005).
3. E. T. Brown, R. Stallard, M. C. Larsen, G. M. Raisbeck, F. Yiu, *Earth Planet. Sci. Lett.* **129**, 193 (1995).
4. F. Von Blanckenburg, *Earth Planet. Sci. Lett.* **242**, 224 (2006).
5. R. C. Fletcher, H. L. Buss, S. L. Brantley, *Earth Planet. Sci. Lett.* **244**, 444 (2006).
6. S. E. Trumbore, C. I. Czimczik, *Science* **321**, 1455 (2008).
7. J. S. Olson, *Ecology* **44**, 322 (1963).
8. K. Van Oost *et al.*, *Science* **318**, 626 (2007).
9. W. Parton *et al.*, *Science* **315**, 361 (2007).
10. M. G. Turner, W. H. Romme, R. H. Gardner, R. V. O'Neill, T. K. Kratz, *Landsc. Ecol.* **8**, 213 (1993).
11. A. F. White, S. L. Brantley, *Chem. Geol.* **202**, 479 (2003).
12. A. F. White, in *Kinetics of Water-Rock Interaction*, S. L. Brantley, J. D. Kubicki, A. F. White, Eds. (Springer, New York, 2008), pp. 463–494.
13. D. A. Wardle, L. R. Walker, R. D. Bardgett, *Science* **305**, 509 (2004); published online 17 June 2004 (10.1126/science.1098778).
14. J. A. Wiens, *Funct. Ecol.* **3**, 385 (1989).
15. O. A. Chadwick, J. Chorover, *Geoderma* **100**, 321 (2001).
16. P. M. Vitousek, H. A. Mooney, J. Lubchenco, J. M. Melillo, J. M. Melillo, *Science* **277**, 494 (1997).
17. G. M. Filippelli, in *Phosphates: Geochemical, Geobiological, and Materials Importance*, M. J. Kohn, J. Rakovan, J. M. Hughes, Eds. (Mineralogical Society of America, Washington, DC, 2002), pp. 391–425.

10.1126/science.1161132

GEOLOGY

An Uncertain Future for Soil Carbon

Susan E. Trumbore and Claudia I. Czimczik

Predictions of how rapidly the large amounts of carbon stored as soil organic matter will respond to warming are highly uncertain (1). Organic matter plays a key role in determining the physical and chemical properties of soils and is a major reservoir for plant nutrients. Understanding how fast organic matter in soils can be built up and lost is thus critical not just for its net effect on the atmospheric CO₂ concentration but for sustaining other soil functions, such as soil fertility, on which societies and ecosystems rely. Recent analytic advances are rapidly improving our understanding of the complex and interacting factors that control the age and form of organic matter in soils, but the processes that destabilize organic matter in response to disturbances (such as warming or land use change) are poorly understood.

There is broad agreement on the major pathways of the soil carbon cycle (see the fig-

ure). Plants are the main source of carbon to soils through tissue residues or via root exudates and symbiotic fungi. These inputs are broken down, transformed, and respired by soil fauna and microorganisms. Some of the carbon converted into microbial biomass and by-products is in turn converted into new microbial biomass (“recycled”) (2). Some organic molecules, such as pyrogenic compounds, may accumulate because of recalcitrance. However, most soil organic matter consists of relatively simple molecules that organize through interactions with surfaces and with each other (3). Organic matter persists in soil mainly because it is physically isolated from decomposition by microbes—for example, by incorporation into aggregates (4) or sorption into mineral (or other organic) surfaces (5, 6). On balance, nearly all the carbon that enters soil as plant residues each year either decomposes and returns to the atmosphere or is leached from soils within a few decades to centuries.

The rates of accumulation and loss of soil carbon are estimated from two kinds of infor-

A detailed knowledge of how carbon cycles through soils is crucial for predicting future atmospheric carbon dioxide concentrations.

mation: direct observations of changes in the amount of organic matter, and inferences based on the age of organic matter as measured by radiocarbon. These rates vary dramatically depending on the time scale of observation, and they reflect differences in the dominant processes contributing to the stabilization of organic matter.

On time scales of months to years, observed rates of mass loss during decomposition of fresh plant litter nearly balance rates of plant litter addition to soils (~2 to 10 Mg C ha⁻¹ year⁻¹). Litter decomposition is thus the major pathway for loss of carbon from soils (see the figure), and rates are controlled by factors such as litter quality, soil faunal and microbial community composition, and climate (7).

On millennial time scales, changes in carbon stocks cannot be observed directly. They are estimated by comparing carbon storage at carefully selected sites that differ in the time since bedrock weathering started (soil age) but are similar in other soil-forming factors such as bedrock material, climate, and vegetation. Such comparisons yield rates of change in soil

Department of Earth System Science, University of California, Irvine, CA 92697, USA. E-mail: setrumbo@uci.edu; czimczik@uci.edu