

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.

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GEOLOGY

An Uncertain Future for Soil Carbon

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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

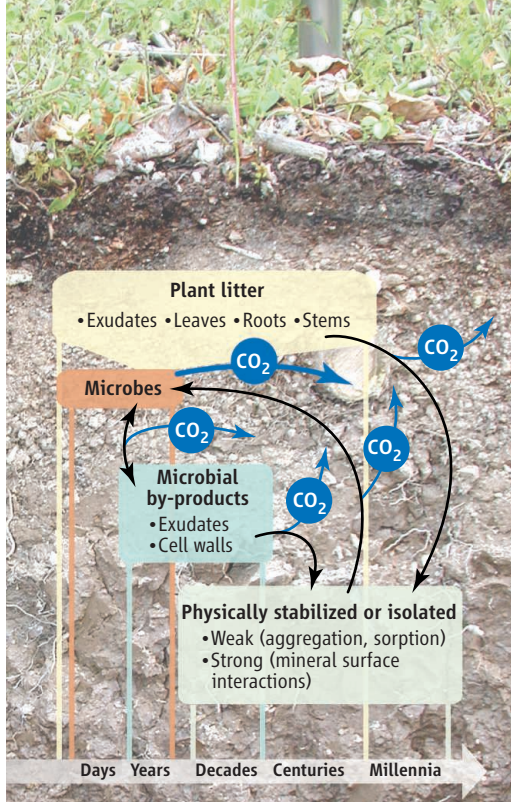
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carbon stores of $\sim 0.02 \text{ Mg C ha}^{-1} \text{ year}^{-1}$, much slower (by a factor of 100 to 500) than fresh litter decomposition (8); on these time scales, the amount and age of soil carbon are controlled by changes in mineral surfaces related to weathering (9).

Most of the concern associated with soil carbon response to global change involves organic carbon stocks that can change over decades to centuries. Changes in these kinds of organic matter are too small to be observed over a few years, and on millennial time scales they are obscured by other factors—such as vegetation productivity and nutrient supply (10)—that vary with soil mineralogy. Our understanding of carbon dynamics on these intermediate time scales relies either on quantifying changes in carbon stocks and stable carbon isotopes after disturbances such as fire or land use change, or on following the incorporation of radiocarbon produced in the 1960s by nuclear weapons tests into soil carbon pools.

Such measurements identify substantial stores of soil carbon that can accumulate and be lost at intermediate rates (~ 0.1 to $10 \text{ Mg C ha}^{-1} \text{ year}^{-1}$). Processes that can stabilize or destabilize organic carbon on these time scales include alterations to the quantity, age, and quality of plant litter inputs; shifts in the makeup, spatial distribution, and function of soil fauna and microbial communities; alteration of weak stabilization processes such as aggregate formation; and changes in mineral surfaces under altered redox or pH conditions. Such processes will respond on decadal time scales to changes not only in climate, but also in nutrient deposition or vegetation. To predict future concentrations of atmospheric CO_2 , it is critical to better understand how much carbon is vulnerable to destabilization on decadal to centennial time scales, and which processes provide the most important controls for a given ecosystem.

The shortcomings of the current understanding are apparent when trying to predict the response of mineral soil carbon stores to global warming. A good example is the debate over the temperature dependence of decomposition rates for different carbon pools (11). In general, fresh plant material decomposes faster at higher temperatures (7). However, it is less clear whether or how carbon stabilized on mineral surfaces responds to temperature changes. Conceptual models like that depicted in the figure are too simplistic in treating production and decomposition as separate rather



Carbon transformation pathways in soil. The scale at the bottom indicates the mean age of organic carbon typically found in each pool. Blue arrows indicate CO_2 production during transformation from one pool to another.

than linked processes (12). Changes in temperature will influence all parts of the soil-plant system; simple temperature functions may provide a means to average across this complexity, but are not likely to be useful for predicting responses outside observed conditions. Progress will require temperature manipulations of whole ecosystems, coupled with observations of soil carbon and isotope fluxes in concert with modeling (13).

The most robust predictions of future soil carbon change involve accelerated decomposition of relatively fresh plant material that persists because of flooding or freezing conditions, rather than by interactions with minerals. For example, high-northern-latitude regions that store vast amounts of carbon in relatively undecomposed forms, and where temperatures are rising faster than the global mean, are predicted to become net carbon sources to the atmosphere over the next century, because decomposition rates increase more than plant productivity does (14).

Outside of these special cases, a number of outstanding issues still limit our ability to predict soil carbon response. For example, plant residues arrive in the soil with different ages (years for leaves, centuries for tree stems). Hence, the radiocarbon age of soil organic matter is not merely a measure of the time period organic matter spends in soils, which may bias interpretations of its stability. Soils are not well-mixed media, and the timing of

degradation and stabilization processes is also regulated by the complex spatial distribution of organic matter, microorganisms, and minerals (5). Sampling that integrates over that spatial domain may mix very young and old components to arrive at an average that is not adequate for describing rates of response to short-term change.

Most detailed studies of soil carbon age and chemistry are conducted on small plots for a few years, yet processes operating at larger spatial scales over decades to centuries (such as erosion, fire, nutrient deposition, or vegetation change) may ultimately determine the impact of soils on atmospheric CO_2 . For example, fire-dominated Mediterranean and boreal ecosystems accumulate surface litter between burning events. Increasing burned area in a given year can return carbon faster to the atmosphere than it accumulates in unburned areas, making the region a net carbon source (15). Rapidly changing land-use patterns, as observed in the tropics, can be more important for evaluating soil carbon balance than are the factors causing variable rates of carbon loss or gain in an individual field (16). Such landscape-scale processes are crucial for the global carbon budget but are only beginning to be addressed in field studies or ecosystem carbon models.

Future progress will come from studies that combine measures of microbial community and activity, soil physics and chemistry, and the structure, age, and chemical nature of organic matter stored in and exiting soils. These studies should not focus only on improving models of the upper 10 to 20 cm of mineral soil at one location, but must recognize that soil processes extend in three dimensions, as deep as roots and across landscapes. Continued changes in climate will ultimately show how soil carbon will respond, but predicting changes would be the safer route given the importance of soil organic matter in sustaining society.

References

1. P. Friedltingstein *et al.*, *J. Clim.* **19**, 3337 (2006).
2. G. Gleixner *et al.*, *Org. Geochem.* **33**, 357 (2002).
3. I. Kögel-Knabner, *Soil Biol. Biochem.* **34**, 139 (2002).
4. J. Six *et al.*, *Plant Soil* **241**, 155 (2002).
5. K. Ekschmitt *et al.*, *Geoderma* **128**, 167 (2005).
6. M. Kleiber *et al.*, *Biogeochemistry* **85**, 9 (2007).
7. W. Parton *et al.*, *Science* **315**, 361 (2007).
8. W. H. Schlesinger, *Nature* **348**, 232 (1990).
9. M. S. Torn *et al.*, *Nature* **389**, 170 (1997).
10. S. L. Brantley, *Science* **321**, 1454 (2008).
11. E. A. Davidson, I. A. Janssens, *Nature* **440**, 165 (2006).
12. M. Heimann, M. Reichstein, *Nature* **451**, 289 (2008).
13. Y. Luo, *Annu. Rev. Ecol. Evol. Syst.* **38**, 683 (2007).
14. K. Dutta *et al.*, *Global Change Biol.* **12**, 2336 (2006).
15. J. W. Harden *et al.*, *Global Change Biol.* **6**, 174 (2000).
16. K. W. Holmes *et al.*, *Biogeochemistry* **74**, 173 (2005).