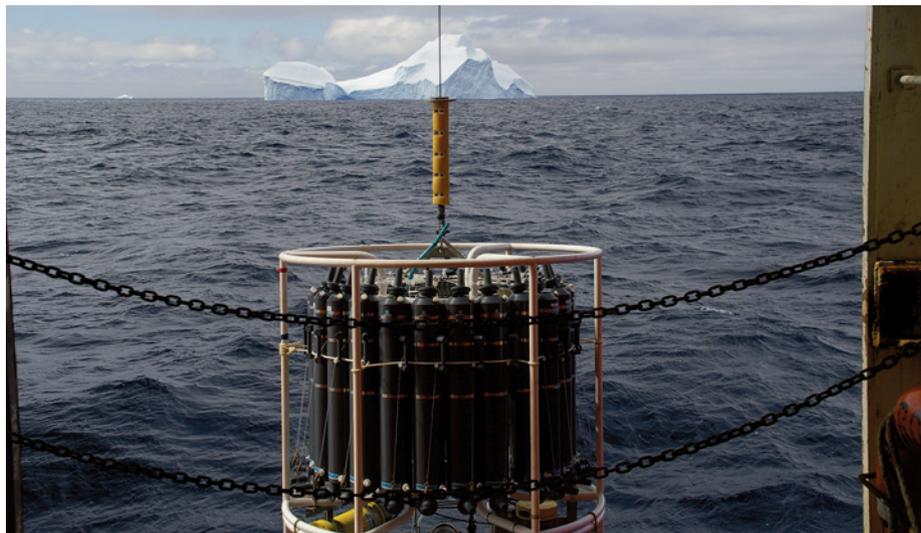


strongly in the Australian Antarctic basin<sup>11</sup>. The largest source of bottom water — the Weddell Sea — has freshened the least<sup>2</sup>. This pattern of change is entirely consistent with glacier melt causing the freshening — the largest sources of melt water are the Pine Island and Thwaites Glaciers<sup>1</sup> located west of the Antarctic Peninsula, and the known circulation patterns would subsequently transport it to the regions of strongest freshening. It is estimated that the fresher AABW can account for about half of the total Antarctic ice-mass loss ( $45\pm 26\%$ )<sup>2</sup>. Although the Southern Ocean and AABW occupy a relatively small fraction of the global ocean, the warming of the water column surrounding Antarctica could account for about  $18\pm 8\%$  of the  $\sim 180$  TW heat gain in the climate system over the late twentieth century<sup>2</sup>; resulting in a rise of the Southern Ocean's sea level of 1–2 mm per year<sup>2</sup>. The accompanying salinity changes also contribute a weak signal to this rise in sea level (as fresher water occupies a greater volume).

Some outstanding issues remain. The freshening AABW accounts for only half of the melt from Antarctic ice-sheet. Surface waters around Antarctica are also freshening<sup>12</sup> and it would seem likely that a significant fraction of the Antarctic melt ends up in these waters. Fresher surface waters are lighter and further isolate sea-ice from the warmer ocean below, and may help explain observed increases in Antarctic sea-ice cover<sup>13</sup>. Closure of the freshwater budget (including changes in rainfall) in the Antarctic region will help in projecting future accumulation and melt of the Antarctic ice-sheet, and therefore our capacity to project Antarctica's contribution to sea-level rise. Although the freshening of AABW is very likely to be coupled with accelerating flow into the sea, and thinning, of the West Antarctic ice sheet, it is less clear whether the melt and subsequent freshening is the result of increased winds in the region



S. R. RINTOUL

Photograph of the standard equipment that is used to observe the ocean from surface to bottom. This instrument delivers the high-quality salinity and temperature measurements needed to monitor changing Antarctic bottom waters. Salinity is measured to an accuracy of better than 0.002 PSS (Practical Salinity Scale) and temperature to 0.002 °C.

or surface warming. Finally, attribution of this AABW freshening to human influences beyond the internal variability of the climate system has yet to be proved. This work<sup>2</sup> provides the initial evidence for exploring these key gaps in our knowledge of the climate system. □

Nathaniel L. Bindoff<sup>2,3</sup> and William R. Hobbs<sup>2,4</sup> are at <sup>1</sup>Antarctic Climate and Ecosystem Cooperative Research Centre University of Tasmania, Sandy Bay, Tasmania 7005, Australia, <sup>2</sup>University of Tasmania, Sandy Bay, Tasmania 7005, Australia, <sup>3</sup>CSIRO Marine and Atmospheric Research, Hobart, Tasmania 7000, Australia, <sup>4</sup>ARC Centre of Excellence in Climate System Science, University of New South Wales, Randwick, New South Wales 2052, Australia.

e-mail: [N.Bindoff@utas.edu.au](mailto:N.Bindoff@utas.edu.au);  
[will.hobbs@utas.edu.au](mailto:will.hobbs@utas.edu.au)

#### References

1. Rignot, E. *et al.* *Nature Geosci.* **1**, 106–110 (2008).
2. Purkey, S. G. & Johnson, G. C. *J. Climate* **26**, 6105–6122 (2013).
3. Church, J. A. *et al.* *Geophys. Res. Lett.* **38**, L18601 (2011).
4. Bindoff, N. L. *et al.* in *IPCC Climate Change 2007: The Physical Science Basis* (eds Solomon, S. *et al.*) 385–432 (Cambridge Univ. Press, 2007).
5. Katsman C. A. & van Oldenborgh G. J. *Geophys. Res. Lett.* **38**, L14610 (2011).
6. Meehl, G. A., Arblaster, J. M., Fasullo, J. T., Hu, A. & Trenberth, K. E. *Nature Clim. Change* **1**, 360–364 (2011).
7. Sokolov, A., Forest, C. & Stone, P. *Clim. Dynam.* **34**, 735–745 (2010).
8. Balmaseda, M. A., Trenberth, K. E. & Källén, E. *Geophys. Res. Lett.* **40**, 1754–1759 (2013).
9. Durack, P., Wijffels, S. & Matear, R. *Science* **336**, 455–458 (2012).
10. Meehl, G. A. *et al.* in *IPCC Climate Change 2007: The Physical Science Basis* (eds Solomon, S. *et al.*) Ch. 10 (Cambridge Univ. Press, 2007).
11. Rintoul, S. R. *Geophys. Res. Lett.* **34**, L06606 (2007).
12. Durack, P. J. & Wijffels, S. E. *J. Clim.* **23**, 4342–4362 (2010).
13. Close, S. E. & H. Goosse. *J. Geophys. Res. Oceans* **118**, 2811–2827 (2013).

## CARBON STORAGE

# A permafrost carbon bomb?

The fate of permafrost soil carbon following thaw depends on hydrology.

Claire C. Treat and Steve Frohling

Lurking beneath Arctic tundra and boreal forest, there is a potential climate-disrupting carbon (C) store in the frozen soil. There is more than twice the amount of atmospheric C in northern permafrost soils<sup>1</sup>,

which are predicted to warm substantially by 2100<sup>2</sup>. In these regions, the mean annual air temperature is generally below freezing, which has resulted in permafrost formation — where the ground/soil remains

frozen throughout the year, except for a thin (typically <1 m) seasonally thawing surface soil, the active layer. Incorporation of organic matter into the permafrost soil limits decomposition of readily available organic



C. TREAT

Permafrost soils, such as those at this tundra site at Imnavait Creek, Alaska, store twice as much carbon as is now in the atmosphere and are expected to thaw due to climate change.

matter and results in large C reservoirs that can be centuries to several thousands of years old<sup>3</sup>. The 1,650 petagram question is what will happen when permafrost thaws and this substantial carbon stock is no longer protected from decomposition by virtue of being frozen? Writing in *Nature Climate Change*, Elberling *et al.*<sup>4</sup> combine field measurements in Eastern Greenland (where continuing permafrost thaw is documented) with lab experiments and modelling to predict trends in C release from thawing permafrost soils.

Whether permafrost will continue to store C with warmer temperatures or release large amounts of soil C is a matter of open debate. Previous studies have shown a range of possibilities: permafrost thaw and warmer soil temperatures may lead to higher soil respiration, resulting in substantial C losses to the atmosphere<sup>5</sup> and release of millennial-age C<sup>6,7</sup>. However, loss of C following permafrost thaw may be slowed by the anaerobic conditions caused by saturation, resulting in lower CO<sub>2</sub> emissions and higher rates of plant productivity and litterfall. This may lead to increased C storage<sup>8</sup>, but also higher methane emissions<sup>9</sup>.

In a continuous permafrost region in Northeast Greenland, Elberling *et al.* measure an increase in the maximum depth of the active layer of 1–2 cm per year from 1996–2008. The field results were compared with a long-term (12-year) experimental

soil incubation in the laboratory for two landforms (drier heath and wetter grass). They found no significant loss of C stocks *in situ* but small to extensive (17–78%) loss during the long-term incubation, with the amount of C loss depending on soil moisture conditions.

Although the authors find large C losses during long-term laboratory incubation, these losses do not translate into *in situ* differences in field C stocks. A comparison of C loss rates from lab incubation results across five similar Arctic sites (ranging from Canada to Svalbard, Norway) confirms that C loss rates from their field-site soils are representative. The difference between field and laboratory results can be explained by looking more closely at the experimental conditions. The incubated soil spent 144 months continuously at 5 °C, while *in situ* soils are thawed (slightly warmer than 0 °C) for less than two months per year. Using a standard ecological temperature sensitivity, Elberling *et al.* estimate that their incubations were equivalent to ~100 years in the field, and the *in situ* exposure is equivalent to ~24 months of incubation. However, much of the incubation carbon loss occurred in the first 24 months, implying that the temperature sensitivity at low temperatures may be much higher<sup>10</sup>. This indicates a need for further incubation studies at near-zero temperatures.

An experimental drainage treatment of the wet grassland site points to a second reason for the field–lab difference. When soils remained very wet, as is common near the surface of permafrost, incubation C losses were low, ~2% loss per year. However, when soils were drained in the laboratory experiment, C losses approached 10–20% per year. These results highlight one of the key uncertainties in the response of soil organic C to permafrost thaw: will ecosystems become drier or wetter — a carbon bomb or slow fizz? Predicting soil moisture following thaws remains a serious challenge. The answer will also influence other climate feedback effects not investigated by Elberling *et al.* — such as vegetation composition<sup>11</sup>, surface albedo and energy balance<sup>12</sup>.

Soil moisture has previously emerged as a control of C losses from other boreal and Arctic permafrost sites<sup>11,13</sup>, but Elberling *et al.* are the first to look at the long-term response of soil C stocks to permafrost thaw in both the field and laboratory, as well as to directly compare results between the two. Moreover, they add a comparison between common Arctic landform types and provide useful information on the sizes of soil C pools and decay rates obtained from the modelling results. However, as the contrasting results between field conditions (no change in C stocks) and incubations (large changes in C stocks) illustrate, there are limitations in extrapolating laboratory results to the field. In addition to issues in soil climate identified by Elberling *et al.*, there are questions about altered microbial communities in incubations<sup>14</sup> and the effects of plant litter and root inputs<sup>15</sup>.

Elberling and colleagues<sup>4</sup> have demonstrated in field, lab and modelling results that permafrost soil organic C is vulnerable to decomposition, but that wet conditions following a thaw can substantially reduce the C loss rate. Therefore the fate of permafrost carbon is closely tied to not only temperature but also hydrology, significantly complicating efforts to predict C losses from permafrost thawing. □

Claire Treat\* and Steve Frolking are at the Earth Systems Research Center, Institute for the Study of Earth, Oceans and Space, University of New Hampshire, Durham, New Hampshire 02834, USA and Department of Earth Sciences, University of New Hampshire, Durham, New Hampshire 03824, USA.

\*e-mail: [Claire.treat@unh.edu](mailto:Claire.treat@unh.edu)

#### References

1. Tarnocai, C. *et al.* *Glob. Biogeochem. Cycles* **23**, GB2023 (2009).
2. Lawrence, D. M., Slater, A. G. & Swenson, S. C. *J. Clim.* **25**, 2207–2225 (2012).
3. Schuur, E. A. G. *et al.* *BioScience* **58**, 701–714 (2008).
4. Elberling, B. *et al.* *Nature Clim. Change* **3**, 890–894 (2013).

5. Natali, S. M. *et al.* *Glob. Change Biol.* **17**, 1394–1407 (2011).  
 6. Dutta, K., Schuur, E. A. G., Neff, J. C. & Zimov, S. A. *Glob. Change Biol.* **12**, 2336–2351 (2006).  
 7. Schuur, E. A. G. *et al.* *Nature* **459**, 556–559 (2009).  
 8. Camill, P., Lynch, J. A., Clark, J. S., Adams, J. B. & Jordan, B. *Ecosystems* **4**, 461–478 (2001).  
 9. Wickland, K.P., Striegl, R.G., Neff, J.C., Sachs, T. *J. Geophys. Res. Biogeosci.* **111**, G02011 (2006).  
 10. Mikan, C. J., Schimel, J. P. & Doyle, A. P. *Soil Biol. Biochem.* **34**, 1785–1795 (2002).  
 11. Johansson, T. *et al.* *Glob. Change Biol.* **12**, 2352–2369 (2006).  
 12. Jorgenson M. T. *et al.* *Can. J. Forest Res.* **40**, 1219–1236 (2010).  
 13. Lee, H., Schuur, E. A. G., Inglett, K. S., Lavoie, M & Chanton, J. P. *Glob. Change Biol.* **18**, 515–527 (2012).  
 14. Sinsabaugh, R. L., Moorhead, D. L. & Linkins, A. E. *Appl. Soil Ecol.* **1**, 97–111 (1994).  
 15. Prater, J. L., Chanton, J. P. & Whiting, G. J. *Glob. Biogeochem. Cycles* **21**, GB4004 (2007).

## SOIL CARBON

# Microbes and global carbon

The amount of organic material that microbes assimilate into their biomass is critical in regulating whether soils, the planet's main pool of organic matter, will absorb or emit carbon in a warmer world.

Joshua Schimel

The Earth system is, to a large extent, a product of microbial activity, yet microbial physiology is almost invisible in computer simulations of the Earth System. Would these models be improved if they treated decomposition as a by-product of microbial growth? Would this improve our ability to predict the planet's carbon cycle and its impact on the climate? Now, writing in *Nature Climate Change*, Wieder *et al.*<sup>1</sup> pose, and attempt to answer these questions.

For most of the Earth's history, the only organisms present were microorganisms (bacteria, fungi and algae). Even now they continue to have a dominant role in the cycling of chemicals around the Earth — the only biogeochemical process carried out predominantly by macroorganisms is photosynthesis. Because plants account for roughly 60% of global primary production, Earth system models (ESMs) increasingly use mechanistic approaches to predict how plant carbon uptake responds to environmental stressors and climate change by explicitly simulating plant physiological processes<sup>2</sup>. So why have we not taken a similar approach to microbial processes?

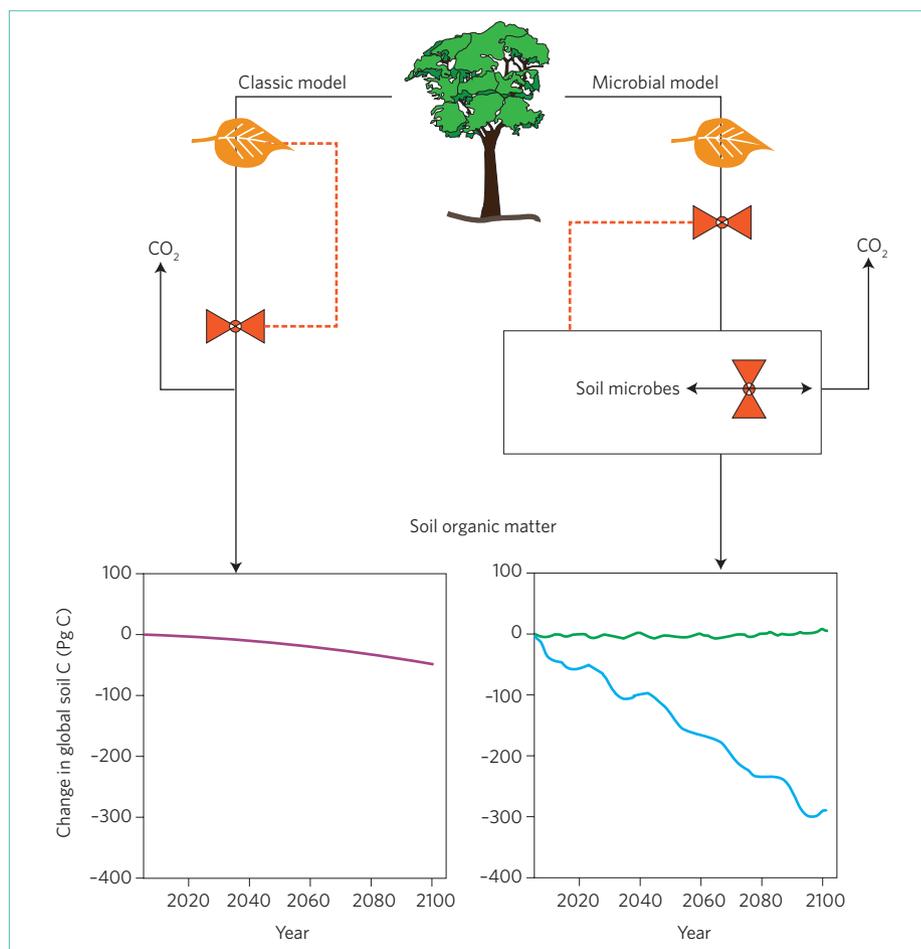
Well, microbial physiology is not absent from ESMs — rather it is an implicit part of their formulation<sup>3</sup>. In ESMs, microbial processes such as decomposition and CO<sub>2</sub> release to the atmosphere are described with first order kinetics (chemical reaction rate) equations. In these, rates are dependent on the size of the substrate pool (the amount of organic carbon present), but are reduced when temperature or moisture are constraining:

$$dC/dt = -k C f_{\text{Temp}} f_{\text{Moisture}}$$

Here  $C$  is the size of the substrate pool,  $k$  is a rate constant that depends on the nature of the substrate and  $f$  is a functional response. There are no microbes visible in

this equation, but if there were no microbes present, the rate constant would be zero.

The microbiology is implicit and microbes' sensitivity to the environment is embedded



**Figure 1** | The stabilization of plant carbon into soil carbon as represented in Earth system models. Left: A classical model, where decomposition is driven by substrate supply. Right: Model by Wieder *et al.* The size of the microbial community drives decomposition, whereas microbial growth efficiency regulates whether carbon is respired to CO<sub>2</sub> or is converted into microbial products. When MGE declines with temperature (green line), microbial biomass and decomposition decline, increasing soil C. When MGE is insensitive to temperature (blue line), biomass increases, accelerating decomposition and reducing soil C. Red symbols indicate major control mechanisms in the models.