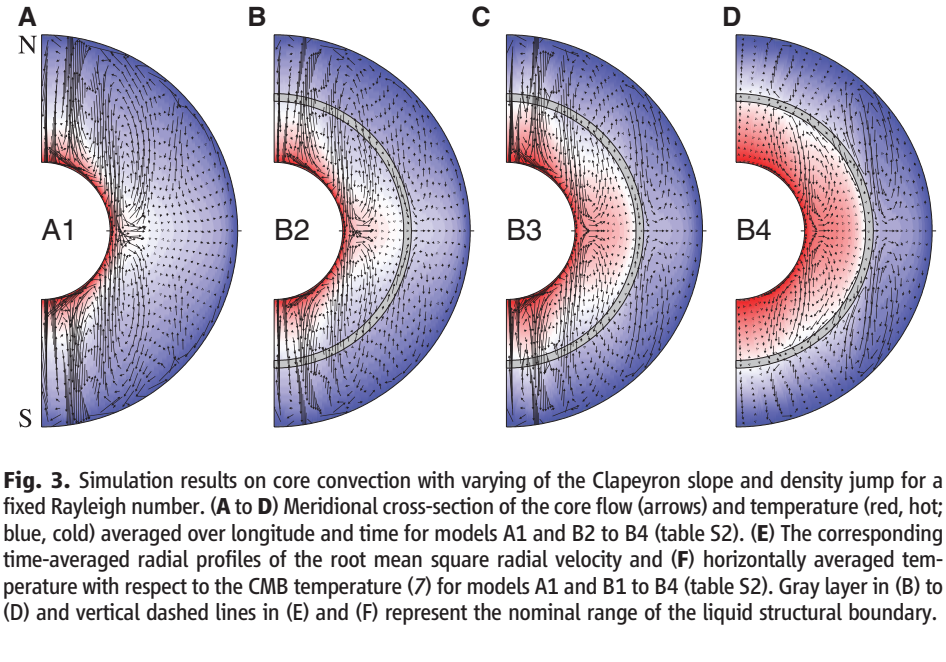

Fig. 3. Simulation results on core convection with varying of the Clapeyron slope and density jump for a fixed Rayleigh number. (**A** to **D**) Meridional cross-section of the core flow (arrows) and temperature (red, hot; blue, cold) averaged over longitude and time for models A1 and B2 to B4 (table S2). (**E**) The corresponding time-averaged radial profiles of the root mean square radial velocity and (**F**) horizontally averaged temperature with respect to the CMB temperature (*7*) for models A1 and B1 to B4 (table S2). Gray layer in (B) to (D) and vertical dashed lines in (E) and (F) represent the nominal range of the liquid structural boundary.

References and Notes

1. F. Birch, *J. Geophys. Res.* **57**, 227 (1952).
2. J. Badro *et al.*, *Earth Planet. Sci. Lett.* **254**, 233 (2007).
3. D. Alfè, M. J. Gillan, G. D. Price, *Earth Planet. Sci. Lett.* **195**, 91 (2002).
4. Y. Fei, H.-K. Mao, *Science* **266**, 1678 (1994).
5. H. Ozawa, K. Hirose, S. Tateno, N. Sata, Y. Ohishi, *Phys. Earth Planet. Inter.* **179**, 157 (2010).
6. R. A. Fischer *et al.*, *Earth Planet. Sci. Lett.* **304**, 496 (2011).
7. Materials and methods are available as supporting material on *Science* Online.
8. R. Jeanloz, *Rev. Mineral.* **14**, 389 (1985).
9. Z. Wu *et al.*, *J. Geophys. Res.* **113** (B6), B06204 (2008).
10. N. Kawai, Y. Inokuti, *Jpn. J. Appl. Phys.* **9**, 31 (1970).
11. J. L. Tallon, *Phys. Lett.* **72A**, 150 (1979).
12. M. Ross, F. J. Rogers, *Phys. Rev. B* **31**, 1463 (1985).
13. S. Urakawa, N. Igawa, O. Shimomura, H. Ohno, in *Properties of Earth and Planetary Materials at High Pressure and Temperature*, M. H. Manghnani, T. Yagi, Eds. (American Geophysical Union, Washington, DC, 1998), pp. 241–248.
14. T. Kinoshita, T. Hattori, T. Narushima, K. Tsuji, *Phys. Rev. B* **72**, 060102 (2005).
15. T. Hattori, T. Kinoshita, T. Narushima, K. Tsuji, Y. Katayama, *Phys. Rev. B* **73**, 054203 (2006).
16. V. V. Brazhkin, S. V. Popova, R. N. Voloshin, *High Press. Res.* **15**, 267 (1997).
17. D. M. Sherman, H. J. F. Jansen, *Geophys. Res. Lett.* **22**, 1001 (1995).
18. N. Sata *et al.*, *J. Geophys. Res.* **115** (B9), B09204 (2010).
19. U. R. Christensen, D. A. Yuen, *J. Geophys. Res.* **90** (B12), 10291 (1985).
20. T. Yanagisawa, Y. Yamagishi, Y. Hamano, D. R. Stegman, *Phys. Earth Planet. Inter.* **183**, 341 (2010).
21. D. Alfè, M. J. Gillan, G. D. Price, *Phys. Rev. B* **65**, 165118 (2002).
22. L. Liu, P. Olson, *Geophys. Res. Lett.* **36**, L10305 (2009).
23. P. Olson, P. Driscoll, H. Amit, *Phys. Earth Planet. Inter.* **173**, 121 (2009).
24. R. A. Fischer, A. J. Campbell, *Am. Mineral.* **95**, 1473 (2010).
25. R. Boehler, *Nature* **363**, 534 (1993).
26. D. Alfè, M. J. Gillan, G. D. Price, *Contemp. Phys.* **48**, 63 (2007).
27. F. D. Stacey, P. M. Davis, *Physics of the Earth* (Cambridge Univ. Press, Cambridge, ed. 4, 2008).

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Supporting Online Material

www.sciencemag.org/cgi/content/full/334/6057/792/DC1
Materials and Methods
Figs. S1 to S10
Tables S1 and S2
References (*28*–*35*)

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Aerosol Indirect Effect on Biogeochemical Cycles and Climate

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The net effect of anthropogenic aerosols on climate is usually considered the sum of the direct radiative effect of anthropogenic aerosols, plus the indirect effect of these aerosols through aerosol-cloud interactions. However, an additional impact of aerosols on a longer time scale is their indirect effect on climate through biogeochemical feedbacks, largely due to changes in the atmospheric concentration of CO_2. Aerosols can affect land and ocean biogeochemical cycles by physical forcing or by adding nutrients and pollutants to ecosystems. The net biogeochemical effect of aerosols is estimated to be equivalent to a radiative forcing of −0.5 ± 0.4 watts per square meter, which suggests that reaching lower carbon targets will be even costlier than previously estimated.

Aerosols are solids or liquids suspended in the atmosphere, and because of the difference in phase, they interact with incoming solar radiation and outgoing planetary radiation (*1*). Because human combustion and land use have substantially increased the amount of aerosols in the atmosphere, the change in radiative forcing from the direct effect of aerosols is estimated to be a substantial cooling (−0.5 ± 0.4 W/m^2) (*1*). In addition, aerosols serve as nuclei for liquid and solid cloud droplets, modifying cloud optical properties, which is termed the aerosol indirect effect. This effect is likely to be as large as the direct radiative effect of changing aerosols, although its magnitude, and even its sign, are highly uncertain (−0.3 to −1.8 W/m^2) (*1*) (Fig. 1). Because the interactions of aerosols with cloud droplets are quite complicated, some authors have argued that the net effect can be positive or negative under different circumstances (*2*, *3*).

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Until recently, an additional climate impact of aerosols has been largely ignored: the indirect effect of aerosols on biogeochemical cycles. Similar to the indirect effect on clouds, these effects occur only to the extent that aerosols affect relevant earth system processes. For example, the indirect effect of aerosols changes cloud droplet properties once clouds form downwind of the aerosol source (cloud albedo effect) and can change the lifetime of the cloud (cloud lifetime effect). Aerosol indirect effects on biogeochemical cycles similarly affect the fluxes downstream from where they are emitted, through one of two different mechanisms: (i) changing the physical climate of the ocean or land ecosystem, and thereby changing biogeochemical fluxes; or (ii) depositing chemicals that modify the biogeochemical cycles. In the latter case, the aerosols could supply either nutrients that stimulate growth or, alternatively, toxins that suppress growth. The aerosol indirect effect on biogeochemical cycles tends to occur on a longer time scale than the aerosol indirect effect on clouds.

The impact of anthropogenic aerosols on biogeochemical cycles through physical processes is just beginning to be recognized. Coupled climate-carbon cycle model simulations suggest that the dominant impact of aerosols on biogeochemical cycles is cooling of the climate (*4*, *5*). Because increasing temperatures tend to decrease the ability of the land and ocean to take up additional carbon (*6*, *7*), this cooling of the planet is likely to allow the land and ocean to take up extra carbon today of an amount between 1 and 14 parts per million (ppm) of CO_2 (*4*, *5*), which translates to −0.02 to −0.24 W/m^2 radiative forcing (*8*). This physical climate forcing is likely to be a combination of the impact of changes in precipitation, temperature, and diffuse radiation, especially on the land carbon cycle (*4*, *9*, *10*).

Once aerosols are removed from the atmosphere, the materials in the aerosols are added to the land or ocean and affect biogeochemical cycles there. On land, the deposition of nitrogen species in aerosols (as well as in gas form) is likely to fertilize many land ecosystems that are nitrogen-limited (*11*–*15*). Estimates of the carbon impact of anthropogenic nitrogen deposition on land range from 0.24 to 0.7 Pg of carbon (PgC)/year, based on observational and model studies (*16*–*20*). Assuming that one-half of the nitrogen deposition derives from aerosols (*21*), we can estimate a radiative forcing of −0.12 to −0.35 W/m^2 from the nitrogen deposition in anthropogenic aerosols [see the supporting online material (SOM) for details of the calculations].

An additional sink of carbon may arise from biomass burning in tropical forests. These forests are likely to be phosphorus-limited (*22*), and biomass burning provides a source of phosphorus to the nonburned vegetation (*23*). Before the recent drought, CO_2 was being taken up by the Amazon at a rate of 0.4 to 1.0 PgC/year (*24*). It is possible that up to one-half of this uptake was due to phosphorus fertilization from biomass burning in the region due to deforestation (which is included in estimates of carbon emissions) (*23*). This would translate to a radiative forcing of 0 to −0.12 W/m^2.

Even though anthropogenic activity has probably caused a large increase in nitrogen and phosphorus deposition to the oceans (*25*, *26*), this is unlikely to affect ocean biogeochemistry significantly because of the large inventories of marine nitrogen and phosphorus (*27*). However, the case for the micronutrient iron is probably different, because iron deficiency is known to limit the growth of phytoplankton in iron-poor regions (*28*), and because nitrogen-fixing organisms require iron to function (*29*). Increases in the deposition of the iron in desert dust since 1870 are likely to have fertilized ocean biota, enhanced nitrogen fixing, and resulted in the uptake of ~4 ppm more CO_2 (*30*), equivalent to a radiative forcing of −0.07 ± 0.07 W/m^2 (*8*).

Aerosol deposition can also be harmful. Many aerosols are acidic (such as sulfates or nitrates), and the deposition of these aerosols onto land ecosystems (called acid rain) can enhance the leaching of nutrients from the system (*31*, *32*). Because some of the acidity that is deposited comes in the form of nitrogen, which also fertilizes the ecosystems (*15*), the magnitude of the impact of acid rain on the land carbon cycle is unclear. Toxic aerosols can also harm ocean ecosystems (*33*), and acid rain can enhance ocean acidification in coastal regions (*34*), but there is no estimate of the magnitude of the effect of these changes on the carbon cycle.

Overall, aerosol indirect effects on biogeochemical fluxes are estimated to be responsible for the extra drawdown of 7 to 50 ppm of CO_2 or a radiative forcing of −0.5 ± 0.4 W/m^2 (*8*), which is similar in magnitude to the direct effects (Fig. 1). The concept of aerosol indirect effect is important because it attributes the changes in CO_2 we observe to the appropriate mechanism. Aerosols in the atmosphere not only are currently counteracting warming from greenhouse gases but, in addition, are enhancing CO_2 uptake by the land and the ocean. There may be additional impacts of aerosols on biogeochemical cycles that affect climate that have not yet been studied [such as the release of other greenhouse gases or aerosols (*35*)].

The identification of additional impacts by aerosols has implications for future climate policy. Emission projections suggest that globally averaged aerosol forcing will decrease (*36*–*42*) as countries reduce emissions to improve their air quality and reduce public health risks. Many aerosols are created in combustion processes at the same time as CO_2 is produced, so reductions in CO_2 emissions could cause corresponding reductions in aerosol emissions. Aerosol effects are

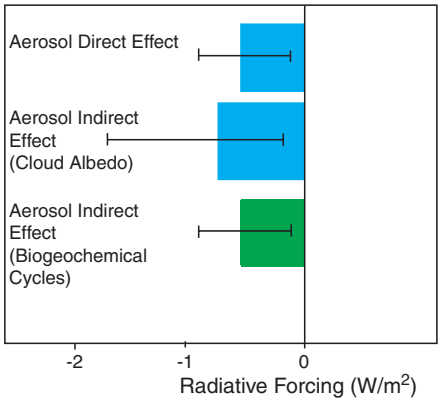

Fig. 1. Aerosol direct and indirect (cloud albedo) radiative forcing estimates (*1*) compared to the radiative forcing estimates from the indirect effect of aerosols from biogeochemical cycles.

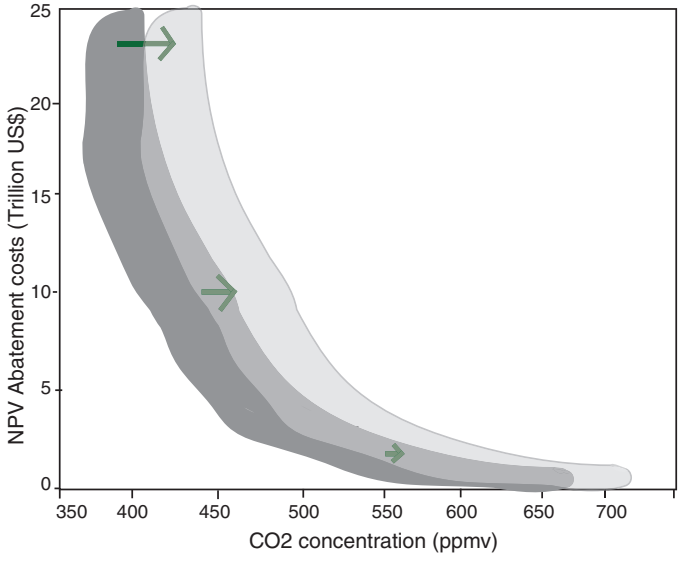

Fig. 2. Net present value (NPV) of abatement costs for different 2100 CO_2 values, based on the range presented in (*36*) (dark gray outline) and shifted by aerosol indirect effect on biogeochemical cycles (green arrows and light gray outline). For each concentration level, the values are shifted by the aerosol indirect effect on biogeochemical cycles, assuming that this effect is equivalent to −30 ppm of CO_2 today (as derived in the text as a range between 7 and 50 ppm) and decreasing this effect as the estimated aerosol radiative forcing decreases in the representative concentration pathways estimated for the next Intergovernmental Panel on Climate Change assessment report (*8*, *36*–*42*).

as diverse as the sources of aerosols, with some aerosols warming the planet (such as black carbon) and others cooling the planet (such as sulfate) (*1*). The net effect of anthropogenic aerosols is to cool the planet (*1*), however, so projected changes in emissions will tend to reduce the aerosol radiative forcing (*36*–*42*). These cuts in aerosols will not only cause an increase in temperatures (*43*) but will also cause a decrease in the uptake of carbon by the land and the ocean. This may preferentially affect the more aggressive carbon policies, because these also will result in the fastest decrease in aerosol emissions (*36*) (SOM). For high–carbon emission cases, changes in emissions should not be large enough to cause significant impacts due to changing aerosol production. However, for the low-emission pathways, even lower emissions will need to be achieved than previously estimated, because of the impact of the aerosol indirect effect on carbon uptake. Although there are many uncertainties in estimating future mitigation costs, it is clear that lower targets for CO_2 concentrations correspond to greater costs. The relationship between cost and targets is highly nonlinear (Fig. 2) (*36*), with costs rising rapidly in a kind of "cliff" as CO_2 targets decrease. Because it is generally not accounted for, the aerosol indirect effect on CO_2 uptake, mediated by biogeochemical cycles as described here, tends to shift the cost cliff toward higher costs for the same CO_2 level at 2100 as compared to when it is ignored (Fig. 2) (SOM). Therefore, achieving lower atmospheric CO_2 concentrations may be even costlier than previously estimated. The estimates provided here suggest that more detailed studies on the effect of aerosols on biogeochemical cycles are important for understanding future climate.

References and Notes

1. P. Forster *et al.*, in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, S. Solomon *et al.*, Eds. (Cambridge Univ. Press, Cambridge, 2007), pp. 130–234.
2. D. Rosenfeld *et al.*, *Science* **321**, 1309 (2008).
3. B. Stevens, G. Feingold, *Nature* **461**, 607 (2009).
4. C. Jones, P. Cox, R. Essery, D. Roberts, M. Woodage, *Geophys. Res. Lett.* **30**, 1479 (2003).
5. N. M. Mahowald *et al.*, *Biogeosciences* **8**, 387 (2011).
6. P. Friedlingstein *et al.*, *J. Clim.* **19**, 3337 (2006).
7. P. Friedlingstein, *Nature* **451**, 297 (2008).
8. In order to convert parts per million of CO_2 to a radiative forcing, we assume a conversion factor of 0.017 W/m^2/ppm of CO_2. See SOM for further details.
9. I. Y. Fung, S. C. Doney, K. Lindsay, J. John, *Proc. Natl. Acad. Sci. U.S.A.* **102**, 11201 (2005).
10. L. M. Mercado *et al.*, *Nature* **458**, 1014 (2009).
11. D. S. LeBauer, K. K. Treseder, *Ecology* **89**, 371 (2008).
12. J. M. Melillo, J. R. Gosz, in *The Major Biogeochemical Cycles and Their Interactions*, B. Bolin, R. B. Cook, Eds. (Wiley, New York, 1983, 2009).
13. F. Magnani *et al.*, *Nature* **447**, 849 (2007).
14. D. Reay, F. Dentener, P. Smith, J. Grace, R. Feely, *Nat. Geosci.* **1**, 430 (2008).
15. R. Q. Thomas, C. D. Canham, K. Weathers, C. Goodale, *Nat. Geosci.* **3**, 13 (2010).
16. E. Holland *et al.*, *J. Geophys. Res.* **102**, 15849 (1997).
17. K. J. Nadelhoffer *et al.*, *Nature* **398**, 145 (1999).
18. A. Townsend, B. H. Braswell, E. Holland, J. Penner, *Ecol. Appl.* **6**, 806 (1996).
19. P. Thornton *et al.*, *Biogeosci. Discuss.* **6**, 3303 (2009).
20. S. Zaehle *et al.*, *Global Biogeochem. Cyc.* **24**, 10.1029/2009GB003521 (2010).
21. Y. Feng, J. Penner, *J. Geophys. Res.* **112**, 10.1029/2005JD006404 (2007).
22. P. Vitousek, *Ecology* **65**, 285 (1984).
23. N. Mahowald *et al.*, *Global Biogeochem. Cyc.* **19**, 10.1029/2005GB002541 (2005).
24. O. Phillips *et al.*, *Science* **323**, 1344 (2009).
25. R. A. Duce *et al.*, *Science* **320**, 893 (2008).
26. N. Mahowald *et al.*, *Global Biogeochem. Cyc.* **22**, 10.1029/2008GB003240 (2008).
27. A. Krishnamurthy, J. K. Moore, N. Mahowald, C. Luo, C. S. Zender, *J. Geophys. Res.* **115**, G01006 (2010).
28. J. Martin, R. M. Gordon, S. E. Fitzwater, *Limnol. Oceanogr.* **36**, 1793 (1991).
29. P. Falkowski, *Nature* **387**, 272 (1997).
30. N. Mahowald *et al.*, *Atmos. Chem. Phys.* **10**, 10875 (2010).
31. G. Likens, *Front. Ecol. Environ.* **8**, e1 (2010).
32. D. W. Schindler *et al.*, *Atmosphere-Biosphere Interations: Toward a Better Understanding of the Ecological Consequences of Fossil Fuel Combustion* (National Academy Press, Washington, DC, 1981).
33. A. Paytan *et al.*, *Proc. Natl. Acad. Sci. U.S.A.* **106**, 4601 (2009).
34. S. C. Doney *et al.*, *Proc. Natl. Acad. Sci. U.S.A.* **104**, 14580 (2007).
35. T. D. Jickells *et al.*, *Science* **308**, 67 (2005).
36. D. van Vuuren *et al.*, *Clim. Change* **81**, 119 (2007).
37. L. Clarke *et al.*, *Scenarios of Greenhouse Gas Emissions and Atmospheric Concentrations. Sub-Report 2.1A of Synthesis And Assessment Product 2.1 by the US Climate Change Science Program and the Subcommittee on Global Change Research* (Office of Biological and Environmental Research, U.S. Department of Energy, Washington, DC, 2007).
38. S. J. Smith, T. Wigley, *Energy J.* **3**, 373 (2006) (special issue).
39. M. Wise *et al.*, *Science* **324**, 1183 (2009).
40. J. Fujino, R. Nair, M. Kainuma, T. Masui, Y. Matuoka, *Energy J.* **3**, 343 (2006) (special issue).
41. Y. Hijoka, Y. Matuoka, H. Hisimoto, M. Masui, M. Kainuma, *Global Environ. Eng.* **13**, 97 (2008).
42. K. Riahi, A. Gruebler, N. Nakicenovic, *Technol. Forecast. Soc. Change* **74**, 887 (2007).
43. M. O. Andreae, C. D. Jones, P. M. Cox, *Nature* **435**, 1187 (2005).

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Supporting Online Material

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SOM Text
Tables S1 and S2

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Recent Synchronous Radiation of a Living Fossil

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Modern survivors of previously more diverse lineages are regarded as living fossils, particularly when characterized by morphological stasis. Cycads are often cited as a classic example, reaching their greatest diversity during the Jurassic–Cretaceous (199.6 to 65.5 million years ago) then dwindling to their present diversity of ~300 species as flowering plants rose to dominance. Using fossil-calibrated molecular phylogenies, we show that cycads underwent a near synchronous global rediversification beginning in the late Miocene, followed by a slowdown toward the Recent. Although the cycad lineage is ancient, our timetrees indicate that living cycad species are not much older than ~12 million years. These data reject the hypothesized role of dinosaurs in generating extant diversity and the designation of today's cycad species as living fossils.

Living fossils and evolutionary relicts are surviving representatives of once diverse or abundant groups. They are noteworthy because they originated tens or even hundreds of millions of years ago yet have persisted with little morphological change. Well-known examples include the coelacanth, the horseshoe crab, the *Ginkgo* tree, and the cycads (Cycadophyta). Fossils indicate the cycads originated before the mid-Permian and reached their peak morphologically, geographically, and in taxic diversity in the Jurassic–Cretaceous (*1*–*4*). Their subsequent decline has been attributed to competition with flowering plants (*5*, *6*) and also to the loss of dinosaurs as dispersal agents (*3*); however, numerical analyses testing a coradiation between dinosaurs and cycads are inconclusive (*7*).

Fossil-calibrated phylogenies (timetrees) were used to test whether living cycads are relics or

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