Climate Change in the Age of Humans

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***Abstract**: The Anthropocene epoch presents a mix of old and new challenges for the world's forests. Climatic instability has typified most of the Cenozoic Era but today's situation is unique due to the presence of billions of humans on the planet. The potential rate and magnitude of future warming driven by continued fossil fuel combustion could be unprecedented during the last 56 million years, and the recovery of atmospheric carbon dioxide concentrations to pre-industrial conditions is likely to last tens of thousands of years. Paleoecological records suggest that responses of forests to human-driven climate change may be complicated by differential mobility and resilience among species as well as by the variable distribution of soil, moisture, and light regimes along latitudinal gradients. The future will be difficult to model and predict precisely, not only due to the inherent complexity of the climate system but also because of uncertainty regarding the dispersal and adaptation of forest species as well as the possible development of new technologies, cultural changes, and the raising of artificial barriers to adaptive migration. Nevertheless, the Anthropocene epoch is a useful concept for re-envisioning modern humankind as a powerful force of nature that will influence the distribution and composition of ecosystems for many millennia to come.*

INTRODUCTION

Human-driven climate change is only one of many challenges that forests must face during the 21st century and beyond. Even without adding more heat-trapping carbon dioxide to the atmosphere than all of the planet's volcanoes combined (Gerlach 2013), the presence of billions of human beings on Earth represents a major source of environmental change. In what is being called the Anthropocene epoch, the Age of Humans, we have become so numerous, our technologies so powerful, and our societies so interconnected that we have become a force of nature on a geological scale.

The Anthropocene term apparently arose spontaneously among many members of the scientific community, including ecologist Eugene Stoermer and chemist Paul Crutzen (Crutzen and Stoermer 2000; Stager 2011), who recognized the scope of human influences in the modern world. There is no consensus yet on when it began. Most definitions date it to the Industrial Revolution, but human impacts on what were previously thought to be "untouched" landscapes have long affected forests through mega-herbivore extinctions, land clearance, fires, and cultivation (Willis and others 2004; Willis and Birks 2006; Lorenzen and others 2011). Although its authorship and timing are difficult to pin down, the Anthropocene concept nevertheless provides a useful context for ecosystem management.

With approximately one quarter of the planet's carbon dioxide reservoir now attributable to our fossil fuel emissions, our behavior has become an integral part of global ecology. Our artificial nitrogen fixation now matches or exceeds natural production of available nitrogen worldwide, we change the appearances of continents through land use practices, rising sea levels, and shrinking ice masses, we disperse some species widely while driving others to extinction, and we direct evolution through changes in gene flow, selective breeding, and genetic engineering. The human presence affects the very survival of forests as well as their distribution, reproduction, and community structure, and it will make the ecological consequences of future climatic changes unique in the history of the planet.

Theoretical modeling provides possible examples of what may lie ahead in terms of climate, but proxy records of geologic history can also help to show which scenarios are most realistic and provide examples of biotic responses to climatic shifts in the past.

CLIMATES OF THE PAST

Today's anthropogenic climatic effects are superimposed on a background of variability that includes both cyclic and irregular fluctuations on multiple spatial and temporal scales. Long, high-resolution records from ice cores, tree rings, cave formations, and aquatic sediments show that abrupt and extreme climate events are not limited to human causes, and that many of today's tree taxa have experienced such changes before.

The last 50 million years of the Cenozoic Era was dominated by cooling from the high-CO_2, hothouse of the Eocene "climatic optimum" (Figure 1). The reasons for this are still unclear, but tectonism, weathering of the continents, and sequestration of carbon in marine sediments are likely contributors to the cooling trend (Garzione 2008) Temperatures fell low enough for an Antarctic ice cap to form between 45 and 34 million years ago, and during the last 3 million years temperatures have dropped far enough to trigger several dozen ice ages.

The overall cooling pattern of the Cenozoic was also punctuated by abrupt warming events. One of the most commonly cited examples was the PETM (Paleocene-Eocene Thermal Maximum) that occurred 56 million years ago and lasted roughly 200,000 years (Figure 1; Dickens 2011). Atmospheric CO_2 concentrations are thought to have reached or exceeded 3000 ppm following the release of 2000-5000 gigatons (Gtons; billions of metric tons) of carbon-rich gases into the atmosphere, possibly through volcanism in the Atlantic basin as well as other factors (Pearson and Palmer 2000; Dickens 2011). Global average temperatures rose by 5-10°C above their already-warm states within 20,000 years or less, plant species migrated poleward, and insect herbivory on foliage increased, possibly in response to higher temperatures (Wing and others 2005; Currano and others 2008). Deciduous redwood forests encircled the Arctic Ocean, Nothofagus beech forests covered Antarctica, and ice-free, richly vegetated continents and land bridges facilitated the rapid migration of species (Bowen and others 2002; Smith and others 2006; Williams and others 2008; Cantrill and Poole 2012).

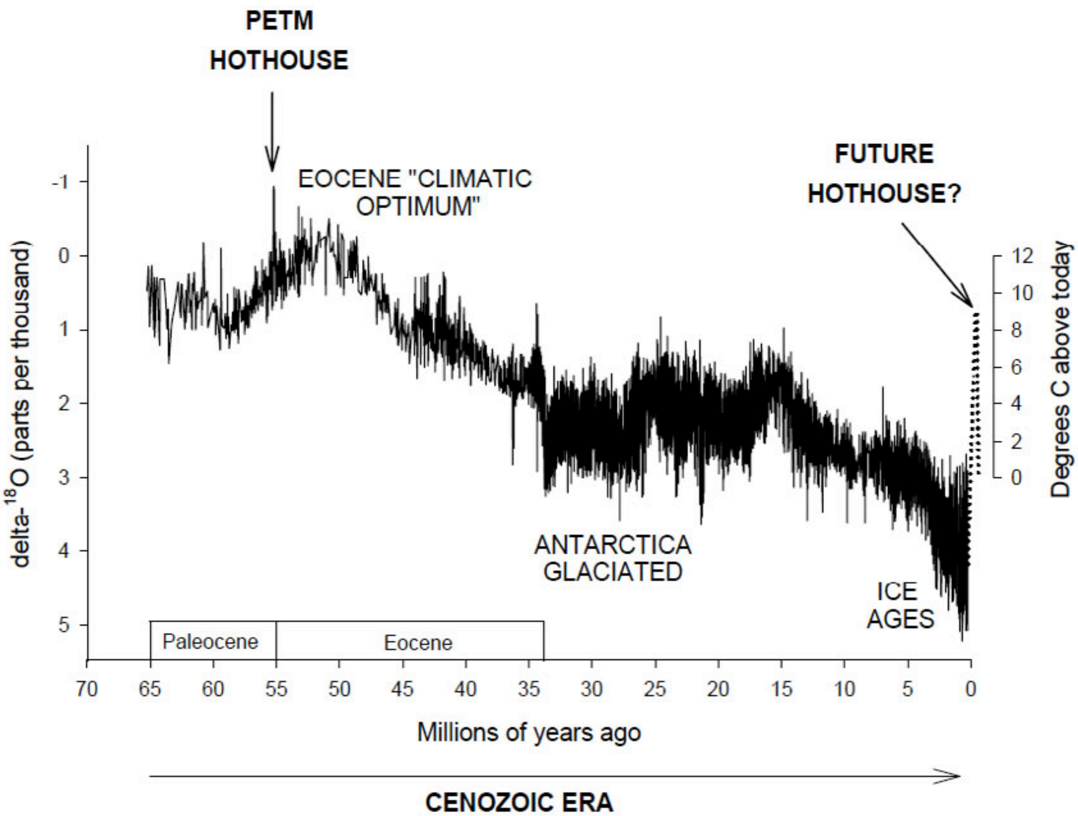

Figure 1. Deep-sea oxygen isotopes and temperatures during the Cenozoic Era (after Zachos and others 2008).

Millennial-scale periodicities in the tilt, wobble, and orbital path of the Earth have been primary pacemakers of ice ages during the last 3 million years. Between cold glacial (longer) and stadial (shorter) periods, seasonal insolation cycles triggered warm interglacials and interstadials, as well. Sediment core evidence suggests that summers became wetter and 8°C or more warmer than today in Arctic Russia during insolation peaks between 3.5 and 2.5 million years ago that included repeated expansion of boreal forest over tundra (Brigham-Grette and others 2013).

The last such warm period, often referred to as the Eemian Interglacial, produced regional temperatures 1-3°C higher than today between 130,000 and 117,000 yr BP (years before present, relative to AD 1950). The Arctic Ocean was partially ice-free but most of the Greenland and Antarctic ice sheets remained intact despite occasional surges that lifted sea levels at least 7 meters higher than today (Blanchon and others 2009; Clark and Huybers 2009; Nørgaard-Pedersen and others 2009). Conifers invaded Siberian tundra north of Lake Baikal, large stands of spruce, pine, and birch developed in southern Greenland, and woodlands in the Adirondack mountains of upstate New York resembled those of today's Blue Ridge, with pollen records from Eemian-age lake deposits revealing the prevalence of oak, hickory, and black gum (Muller and others 1993; De Vernal and Hillaire-Marcel 2003; Granoszewski and others 2004). Rainfall intensified abruptly over 200 years or less in monsoonal Asia, and greener, moister conditions in tropical Africa and the Middle East helped Stone Age peoples to migrate through what are now the Sinai and Negev deserts (Schneider and others 1997; Chen and others 2003; Yuan and others 2004; Vaks and others 2007).

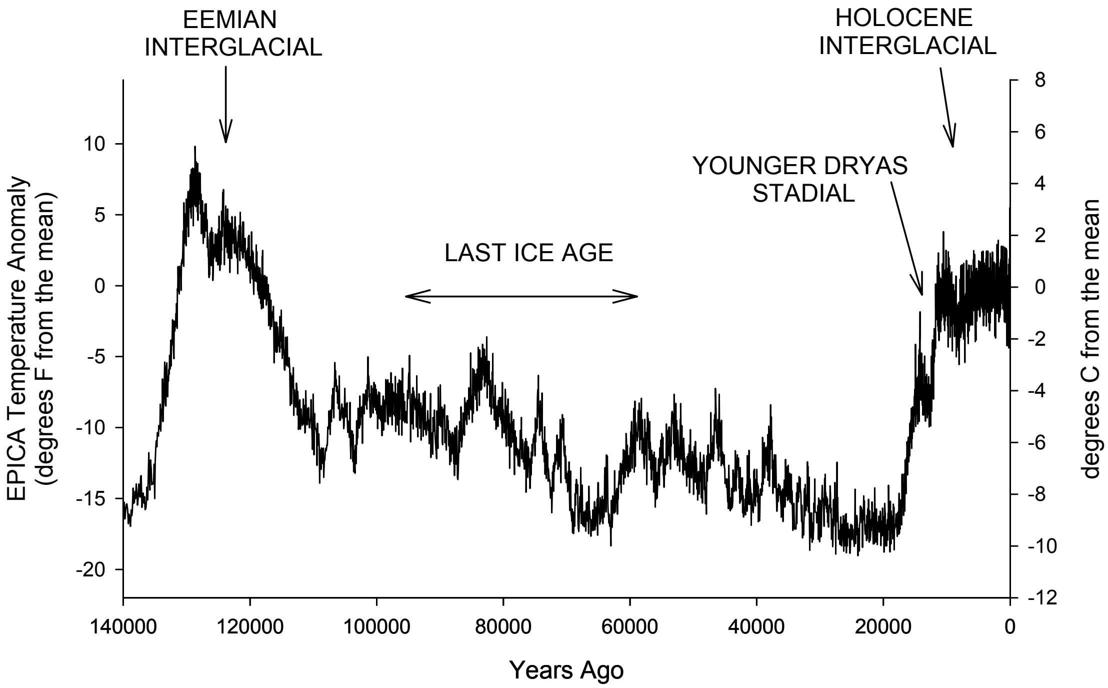

Figure 2. EPICA ice core record of Antarctic temperatures from the Eemian Interglacial to present (after EPICA Community Members 2004).

More rapid and short-lived disruptions also occurred during glacials (Figure 2), including Dansgaard-Oeschger cycles and Heinrich events associated with ice sheet surges and extreme climate fluctuations. Around 17,000 yr BP, massive ice-rafting and cooling in the North Atlantic basin contributed to a sudden, catastrophic collapse of the Afro-Asian monsoon system which desiccated Lakes Victoria, Tana, and Van, and produced genetic bottlenecks in human populations in India ("Heinrich Stadial 1;" Stager and others 2011). Around 13,000 yr BP, the Younger Dryas stadial represented an abrupt return to glacial-type conditions in much of the northern hemisphere that began within less than a decade in some locations and caused severe aridity in much of tropical Africa and southern Asia (Mayewski and others 1993; Stager and others 2002). The end of the Younger Dryas 11,700 years ago represented a rapid shift to the warmer conditions that have dominated the Holocene epoch to modern times.

During the last 11,700 years, the fluctuations preserved in ice core records were not as dramatic as they were during the preceding glacial period, leading to a common misperception that climates of the Holocene were stable before the Industrial Revolution. In fact, ecologically significant instability was still common, even at the poles (O'Brien and others 1995; Mayewski and others 2004). High summer insolation in the northern hemisphere during the early Holocene contributed to ice retreat on the Arctic Ocean and the expansion of lakes and forests throughout tropical Africa (DeMenocal and others 2000; Stager and others 2003; Kaufman and others 2004), the effects of El Niño-Southern Oscillation (ENSO) increased notably after about 5000 yr BP (Moy and others 2002), and other rapid climate changes also occurred throughout the Holocene (Mayewski and others 2004).

Within the last millennium, regional warming during the Medieval Climate Anomaly (ca. 1000-700 yr BP) brought severe drought to much of North America and East Africa and more widespread cooling occurred during the Little Ice Age (ca. 600-200 yr BP) triggering alpine glacial advances in Europe (Mayewski and others 2004; Maasch and others 2005; Verschuren and others 2009). During the last century, non-human sources of variability including ENSO, the North Atlantic Oscillation, shifting westerly wind tracks, and the eleven-year solar cycle have repeatedly disturbed temperature and precipitation regimes over wide areas of the planet (IPCC 2007; Stager and others 2007, 2012).

In sum, high-resolution paleoclimate records reveal far more natural variability than was once assumed from earlier work that failed to sample geological archives in sufficient detail. The rapidity of recent climatic changes is not, as has sometimes been suggested, by itself sufficient evidence to identify humans as the cause.

The ancestors of today's forests experienced numerous climatic shifts in the past, so change alone is not a unique threat in and of itself. However, these records also offer stern warnings about what may lie ahead as a result of human activities in the Anthropocene. The idea of an ice-free Earth, acidified oceans, and massive, rapid climatic disruptions due to greenhouse gas buildups are not merely fantasies among doom-and-gloom radicals; we now know that such things really can happen because they have happened before, even without major human impacts. And although a facile interpretation of geological history might lead one to conclude that climate change is not a threat because it is "natural and ongoing," a more careful reading of paleoecological records shows that extreme climatic shifts of the past would be most unwelcome in today's world with more than 7 billion human beings in the picture.

CLIMATES OF THE FUTURE: THE LONG TERM

What does the future hold? Climates will continue to change as they always have, but the effects of human presence will redefine the baselines upon which natural variability plays out. We are essentially loading the world's weather dice through a hotter, more vigorously circulating atmosphere. The limitations of models along with uncertainties regarding human behavior and technology forbid precise portrayals of what lies ahead, but the general direction and nature of global-scale changes are clear. The more greenhouse gases that we release, the higher global mean temperatures will become and the farther inland oceans will advance. Paleoclimate records also show that, in general, large-scale warming has tended to increase the water content and extent of monsoon systems, to shift mid-latitude storm tracks poleward, and to reduce the extent of ice sheets, glaciers, and sea ice.

Surprises can also emerge from such a complex system, however. For example, although much of tropical Africa became more arid during northern hemisphere coolings and tends to experience more intense rainfall in years just prior to solar maxima, an as-yet unexplained reversal of the cool-dry, warm-wet relationship produced dramatic lake level rises in East Africa during a prolonged solar minimum of the cool Little Ice Age, thereby weakening confidence in our understanding of how tropical climates operate (Stager and others 2005, 2007; Verschuren and others 2009). It has also been proposed that retreat of Arctic sea ice during recent years has contributed to rapid, erratic, and extreme swings in regional climates of the northern hemisphere (Francis and Vavrus 2012). We will not be able to model our way into complete preparedness for everything that the climate system may do in the future, but reasonable generalizations can nonetheless be made from long-term perspectives on the nature and causes of climate change.

One source of insights into future carbon dioxide dynamics is the work of scientists such as David Archer, whose pioneering research at the University of Chicago has been corroborated by other investigators as well (Archer 2005; Archer and Brovkin 2008; Eby and others 2008; Schmittner and others 2008). The astoundingly long-term views of the future that these studies provide show that we are setting in motion a much larger and longer-lasting array of disruptions than the relatively short-term global temperature rise that currently occupies our attention (Stager 2011).

At the heart of these findings is a simple question: "where does carbon dioxide go when it leaves our smokestacks and tailpipes?" Roughly three quarters of it will dissolve directly into the oceans during the next several centuries to millennia, leaving slow weathering of carbonate and silicate minerals to wash the airborne remainder into the sea over tens of thousands of years (Figure 3). When fossil fuel emissions inevitably level off and decline, whether by design or by depletion, marine uptake will cause atmospheric CO_2 concentrations to level off and then to drop nearly as steeply as they rose until the oceans can absorb no more and mineral reactions more slowly consume the leftovers. During the relatively brief turnaround phase of "climate whiplash," many of the selection pressures that operated in the context of rising temperatures may swing into reverse during the cooling that follows (Stager 2011).

The form and timing of the peak, whiplash, and the long tail of the cooling-recovery curve will largely depend upon how much carbon dioxide we release during the next century or so. In a relatively moderate emissions scenario such as B1 (IPCC 2007) in which non-fossil energy sources quickly replace coal, oil, and gas, approximately1000 gigatons (Gtons) of carbon will have been emitted since the Industrial Revolution. If instead we burn all remaining fossil fuel reserves in a scenario more like A2 (IPCC 2007), then a total discharge of closer to 5000 Gtons is more likely. This would lead to a higher, later, and more protracted peak and a much longer recovery (Figure 3).

In one moderate scenario in which emissions decline after AD 2050, atmospheric concentrations of CO_2 reach 550-600 ppm by AD 2200 (Figure 3; Stager 2011). At thermal maximum around AD 2200-2300, global average temperatures are 2-4°C higher than today. After a whiplash stage lasting several centuries, CO_2 concentrations decrease steeply for several millennia due to marine uptake, and then fall within the range of pre-industrial conditions after tens of millennia, possibly as long as 100,000 years. Even in this relatively mild case, the thermal effects of the excess carbon dioxide in the atmosphere are likely to prevent the next ice age, which orbital cycles could otherwise trigger around AD 50,000 (Berger and Loutre 2002; Archer and Ganopolski 2005).

In a more extreme scenario, CO_2 concentrations peak close to 2000 ppm around AD 2300 and take at least 400,000 years to recover (Figures 3,4; Stager 2011). The whiplash stage lasts for several thousand years, producing a seemingly stable plateau of PETM-style warmth 5-9°C warmer than today that could persist long enough for ecosystems and cultures to co-evolve with before the long recovery period destabilizes them again.

In both scenarios, the staggered responses of temperature and sea level to changing CO_2 concentrations further complicate environmental settings for future forests as well as for human beings. In Figure 4, for example, global mean temperature continues to climb for several centuries after the CO_2 peak, and sea levels continue to rise for thousands of years after the thermal peak because the temperatures are still high enough to melt continental ice masses.

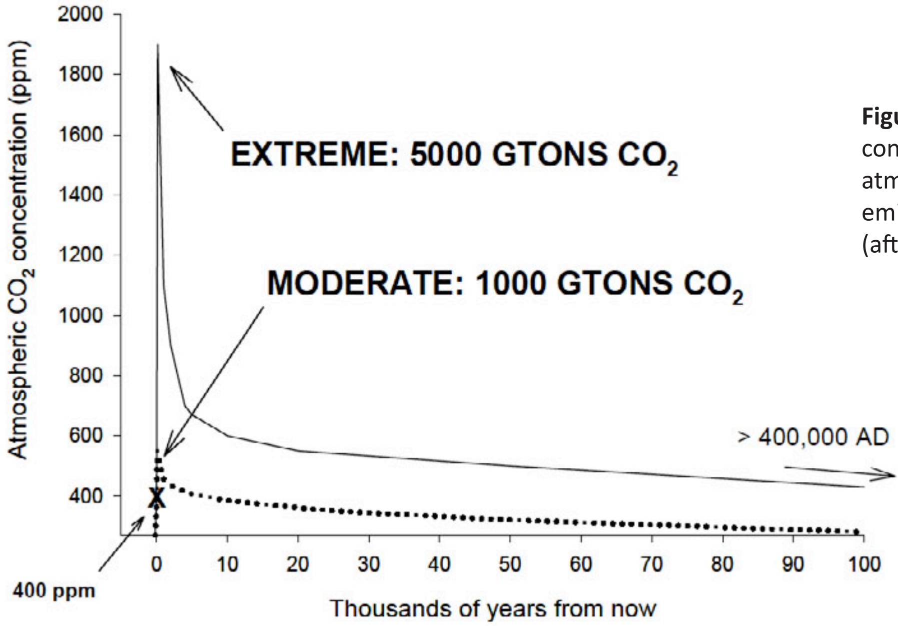

Figure 3. Carbon dioxide concentrations in the atmosphere under two emissions scenarios (after Archer 2005).

What does the geological record reveal about possible consequences of such scenarios for future forests? The more moderate case has much in common with the Eemian Interglacial. Although it was not caused by greenhouse gas buildups, it did produce conditions warmer than today in many locations, particularly during summers in the northern hemisphere. Even though it lasted for about 13,000 years, it failed to de-ice the planet entirely, and polar bears and other arctic biota survived it. Extensive poleward migrations of forests and animals resulted, and rapid changes in sea level followed sporadic destabilizations of ice sheets, eventually submerging much of Florida.

The more extreme case has much in common with the PETM, which did result from a greenhouse gas release comparable to our own. Fossil carbon is depleted in the stable isotope, ^{13}C, and a dramatic global decline in delta-^{13}C in PETM sediments due to enormous geological inputs of fossil carbon into the air and oceans is a diagnostic marker for that event. A similar global dilution of ^{13}C content of the world and its inhabitants is currently under way as a result of our own fossil carbon emissions, and its isotopic signal is being preserved in the geologic record as an anthropogenic sequel to the PETM. Warming during the PETM increased the intensity of rainfall, weathering, and runoff over most of the planet, and it left no refuge for cold habitats.

Conditions similar to those of the PETM are likely to develop again in a "burn-it-all" emissions scenario, but several factors will differ in an Anthropocene reprise of former hothouse states. A modern return to the CO_2 concentrations and temperatures of 56 million years ago would involve increasing those parameters from a much lower thermal baseline that currently allows extensive cold-dependent ecosystems to exist at high latitudes and altitudes. The overall pace of the changes associated with such a return to an ice-free world, were they to occur over a span of several centuries, would outstrip those of the PETM and similar greenhouse warmings of the earlier Cenozoic which occurred when the atmosphere and oceans were already warmer than now (Figure 1).

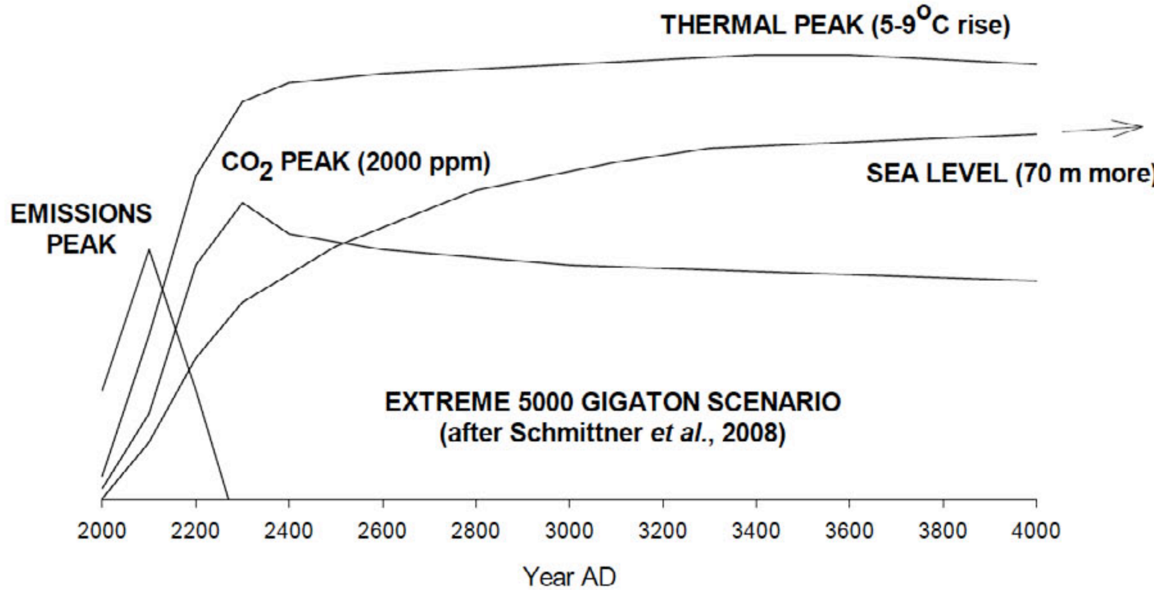

Figure 4. Sequential environmental changes expected in an extreme 5000 Gton carbon emission scenario (after Schmittner and others 2008).

The responses of forests so far back in time may not be directly comparable to those of the Anthropocene, but more recent sediment records suggest that many of today's plant species may be quite resilient to climatic fluctuations if they are free to migrate in response. Pollen and other data from Arctic Russia show that when summer temperatures there were 8°C or more higher than today 3.5-2.5 million years ago, forests in the region changed their geographical distributions and abundances but still consisted of larch, pine, birch, alder, spruce, and other taxa that have also persisted through multiple glacial and interglacial periods to the present day (Brigham-Grette and others 2013). Human presence, however, could seriously restrict such adaptive movements during the Anthropocene.

We cannot know exactly what new technologies will eventually arise or how future societies will respond to the climatic settings that we bequeath to them. Perhaps an ice-free Arctic will come to seem both natural and preferable to the frozen state that we now consider to be normal, and what we would call "recovery" might be experienced as a global cooling disaster thousands of years from now. Even so, potentially important insights arise from such long-term perspectives:

(1) Human influences on the planet have become more powerful than many of us yet realize.

(2) Rapidity of climatic change can be more ecologically stressful than the magnitude or direction of change. The last century's warming (ca. 0.7°C) proceeded at least twice as quickly as the onset of the Eemian Interglacial, and in an extreme emissions scenario global mean temperature could rise by 2-5°C per century between now and AD 2300 (Figure 4), significantly faster than the onset of the PETM.

(3) Although extreme climatic instability has occurred before, the restriction of free migration and other human impacts now make such instability more challenging for species and ecosystems of the Anthropocene.

CLIMATES OF THE FUTURE: THE 21ST CENTURY

Although the upward direction of global average temperature change is easy to anticipate, variable responses within different components of the climate system will make accurate prediction of regional and local-scale conditions difficult. Long-term, global-scale climates are easier to simulate and predict than the more here-and-now, down-to-earth scales of change that many forest managers and urban planners deal with. The inherent limitations of global climate models can be magnified when they are downscaled to focus on relatively short time scales and specific regions, and demand for detailed projections on the regional and local scales sometimes leads people to ask more of climate models than they can reliably produce (Hulme and others 2009; Hefferman 2010; Schiermeier 2010; Trenberth 2010). The limitations of global climate models are often magnified when they are downscaled to focus on relatively short time scales and specific regions, and linking these in turn to models of hydrology or biological processes can amplify errors further (Schiermeier 2010; Trenberth 2010; Beier and others 2012).

A recent comparison of 16 commonly cited models that were downscaled to the Lake Champlain watershed of Vermont and New York illustrates some of the problems that may be faced in such studies (Stager and Thill 2010). All of the models anticipated significant warming by AD 2100, but they disagreed on the magnitudes and seasonality of the changes. The question of seasonality has serious implications for forest ecology because the distribution of temperatures through the year affects ecologically important factors such as snowpack, spring runoff, and net water balance in summer. Perhaps the most reliable projection regarding future temperature in this region may simply be that it will increase as greenhouse gas concentrations rise, which one could conclude even without the aid of models.

Precipitation patterns are also important to forests, but they are more difficult to simulate and to predict (Schiermeier 2010). Precipitation can vary tremendously over small geographical areas, making it difficult to obtain accurate observational records of regional precipitation alone, much less to develop accurate predictive models. Topography, humidity, wind direction and speed, albedo, and other factors further complicate regional-scale modeling of future precipitation. ENSO also strongly influences precipitation patterns around the world, but there is as yet no consensus regarding its likely behavior in the future, and similar uncertainty obscures the future of the North Atlantic Oscillation, Pacific Decadal Oscillation, and other sources of variability, as well (IPCC 2007). In the case of the Lake Champlain study, the 16 regional precipitation scenarios varied in magnitude, seasonality, and even in the direction of trends. Although it is common in such cases to note that the ensemble average of the models states thus and so, it is difficult to know in advance which models are the most accurate, and sticking with the majority may mean rejecting a more reliable minority.

Despite these limitations, observational data and paleoclimate records can document regional and local patterns that have accompanied global-scale changes of the last century, and which can help to inform speculations about future changes. Such data indicate that poleward retreat of the austral westerly wind belt in a warming future could reduce winter rainfall over southwestern Australia and the fynbos region of South Africa (Biastoch and others 2009; Stager and others 2012). A warmer atmosphere is likely to be more energetic and turbulent and to carry more water vapor from the oceans and vegetation (IPCC 2007), and an associated widening of the tropical rain belt has already been observed (Seidel and Randel 2007). The effects of such processes are also apparent in historical records of decreasing ice cover on Lake Champlain and a recent rise in the intensity of extreme precipitation events which, in turn, supports model projections of similar trends in a warmer future (Figure 5; Stager and Thill 2010).

Even if climate models cannot tell us exactly what will happen in the future with complete certainty, they can still provide valuable examples of what realistically could happen. High-quality, multi-parameter simulations can reveal unexpected consequences from perturbations in complex systems that might otherwise be overlooked, and they provide ballpark ranges of variability that can sometimes be refined further through reference to historical records. The spectrum of precipitation variability suggested by diverse arrays of models may also provide helpful estimates of the possible magnitudes of such changes.

As noted in the case of the Champlain watershed, most downscaled models anticipate warmer and wetter conditions by AD 2100, but some suggest that aridity may prevail in some seasons as well (Stager and Thill 2010). Preliminary results from the analysis of fossil diatoms in lake sediment cores indicate that severe droughts occurred in the Champlain basin during the Medieval Climate Anomaly, which links warming to aridity as the minority of models do. However, extreme droughts also occurred during the subsequent cool Little Ice Age, and local observational records show higher lake levels and rising storm intensities have accompanied warming since the 1960s. From what at first appears to be confusion over the future of precipitation in this region, a useful insight emerges: we cannot expect precipitation patterns in the watershed to remain stable, nor can we expect them to be entirely predictable. This, in turn, highlights the need to build resilience into forest management plans.

FOREST MANAGEMENT IN THE ANTHROPOCENE

Forests have long experienced climatic instability in the past, but the most extreme, global-scale disruptions were uncommon in relation to the lifetimes of individual organisms. In addition, most of those events were not accompanied by very large increases in atmospheric carbon dioxide concentrations: Eemian CO_2 levels, for example, remained well below the 400 ppm that prevails at the time of this writing. Perhaps most importantly, none of the extreme climatic shifts of the past occurred while such a large array of additional anthropogenic factors were in operation.

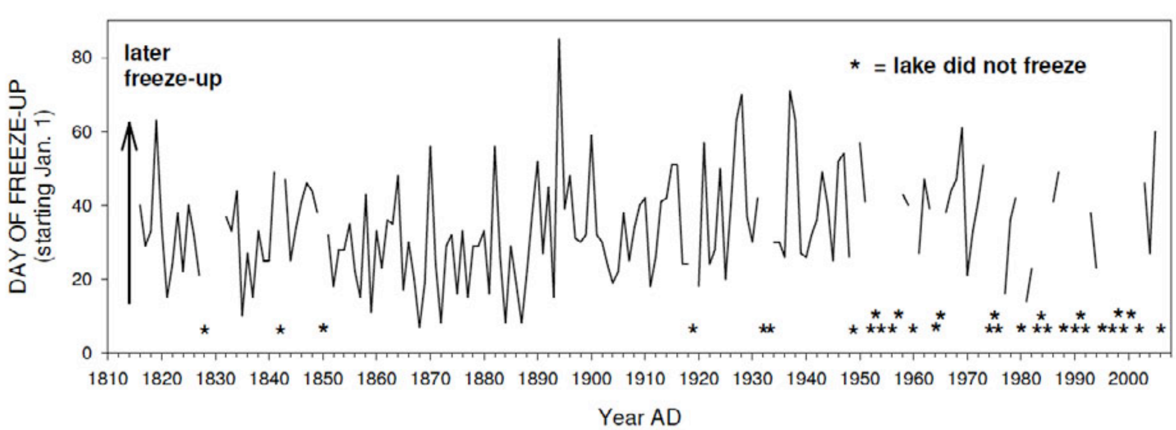

Figure 5. Freeze-up dates for the main body of Lake Champlain. Asterisks indicate winters in which the lake did not completely freeze (after Stager and Thill 2010).

We are sailing into an uncertain future with no exact historical analogs to inform us and little more than educated guesswork to guide us, but it is clear that the rapid pace and global extent of Anthropocene warming will make it increasingly difficult to preserve biotic communities in their present form.

The ecological effects of long-term climatic change may be particularly abrupt on regional and local scales when physiological or physical thresholds are crossed. When winter temperatures are no longer cold enough to exclude an herbivorous insect or pathogen from a habitat, sudden changes in forest composition may result (Dale and others 2001), and the melting point of snow and ice is a fixed boundary across which today's seasonally cold ecosystems will be pulled as the world warms. Old-growth forests that were established in western North America under cool, moist conditions of the Little Ice Age may become increasingly vulnerable to sudden replacement by other taxa through future disturbance events (Willis and Birks 2006). In addition, latitudinal shifts and meanders in major wind belts can bring rapid, extreme climatic changes to sites that lie adjacent to or beneath them.

Paleoecological records show that vegetational communities did not always move as coherent units during climatic shifts of the past, and differential migration rates and climatic tolerances of different species will likely produce new combinations in a warming future (Pitelka and others 1997; Williams and Jackson 2007). As paleoecologist Tom Webb (1988) once said, "plant assemblages are to the biosphere what clouds, fronts, and storms are to the atmosphere... they are features that come and go." Current distributions of tree species do not necessarily reflect their true potential bioclimatic niches, which may further limit our ability to predict their responses to future changes, and the effects of climatic shifts on animal communities can indirectly influence forests as well. The recent lack of winter ice on Lake Superior, for example, may be helping to alter forest structure on Isle Royale because it more fully isolates the local wolf population from the mainland. The negative effects of inbreeding among the wolves, in turn, encourage resident moose populations to expand and more heavily browse the island's forests (Mlot 2013).

As warming pushes isotherms poleward in coming centuries, they may force taxa to enter regions in which their preferred soils, precipitation patterns, and/or light regimes are not available, and no-analog communities and novel environmental settings are likely to emerge in the future as they have in the geological past (Williams and Jackson 2007). If rising temperatures open the high Arctic to colonization by forests, for instance, plants that can tolerate the long darkness of circumpolar winters will be most likely to replace what is now tundra, potentially creating vegetational assemblages that have not existed since the early to mid-Cenozoic or, possibly, at any previous time in history (Sturm and others 2003).

Although human intervention may produce new or "unnatural" ecosystems in the future, paleoecological records show that novelty is not unusual in Earth history. Humans have been affecting forest composition for tens of thousands of years, and differential rates and modes of dispersal in the face of environmental instability have often led to new combinations of species. Some of the vegetational communities that we are familiar with today are surprisingly young; oak savannas of northeastern Iowa and grassy montane parklands of northwestern Wyoming originated roughly 3000 years ago, ponderosa pine forests of the Bighorn Basin arose 2000 years ago, and mixed northern hardwood/conifer forests of northeastern Michigan may be only a thousand years old (Jackson 2006 2012). In light of the ephemeral nature of many plant communities and the huge extent of the modern human footprint, a long-term Anthropocene perspective suggests that managing ecosystems in ways that include difficult trade-offs, triage, or new combinations of species does not necessarily make them too unusual or unnatural to be desirable (Kareiva and others 2007).

Human behavior will probably have the largest but least predictable influences on the future of forests in the Anthropocene. The amount of heat-trapping carbon emissions that we eventually release will determine the trajectory of climatic change for tens of thousands of years to come. Our definitions of desirable and undesirable species and ecological processes are likely to change over time, with enormous consequences for the future of biodiversity. And in a world where the human presence can both aid and hinder the adaptive migration of species, our ability to help organisms to colonize new settings will become increasingly critical to their survival.

The unsteady nature of future climate will make the long-term stability of many forest communities an impossible goal, particularly for those that are confined within static and/or shrinking borders. Flexibility and resilience in the face of environmental and cultural change will become ever more important to the sound management of ecosystems in this new and unprecedented Age of Humans.

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