An Economist's Guide to Climate Change Science

Solomon Hsiang and Robert E. Kopp

Humans have engaged in large-scale transformation of natural systems for millennia. Stone Age hunting technologies led to extinctions of large mammals; agricultural revolutions transformed forests into farmlands; pursuit of minerals has carved the earth's surface; dams and reservoirs now manipulate the flow of almost all rivers; and synthetic fertilizers now flood the nitrogen cycle. But among these transformations, the restructuring of the global carbon cycle and the accompanying alteration of the climate stands apart in its sheer scale, complexity, and economic significance. Essentially all humans that have ever lived contributed, in their own small ways, to reshaping this planetary-scale system. Thousands of years of forest clearance may have added hundreds of billions of tons of carbon to the atmosphere. In the industrial era, every home lit by a coal or natural gas-fired power plant and every petroleum-powered train, plane, and motor vehicle has contributed to the net accumulation of carbon dioxide in the atmosphere. The average human contributes about 5 tonnes of carbon dioxide (CO_2) every year (Le Quéré et al. 2018), about a quarter of which will remain in the atmosphere for well over a millennium (Archer et al. 2009).

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Those emissions of CO_2, together with other greenhouse gases, distort the planet's energy balance. In steady state, the sunlight that makes it to the Earth's surface is absorbed and then re-radiated to space as an equal quantity of heat (technically, infrared light). The accumulation of greenhouse gases in the atmosphere blocks some of this re-radiation, redirecting energy back toward the Earth's surface: about 27 trillion watts (0.05 watts per square meter) per 1 percent increase in atmospheric CO_2 concentrations, equivalent to the energy of one Hiroshima-scale atomic bomb spread over the surface of the Earth every 2.3 seconds. The resulting climatic distortion affects not just temperatures around the world, but also where clouds form, when it floods, how cyclones move, and the volume of water in the ocean. Thus, while fossil-fueled human industriousness has raised unprecedented multitudes out of poverty, the scale of the climate change externality it has produced is similarly extraordinary.

At least since Nordhaus's (1977) presentation at an American Economic Association annual meeting, the analysis and management of climate change has been recognized as an important economic problem, and a growing number of economists are lending the world their expertise in understanding the problem and developing solutions. However, conversations with colleagues indicate to us that a general discomfort with physical sciences—a subject sometimes not studied since high school—prevents many economic minds from engaging more deeply with the problem of climate change.

The goal of this article is to provide a brief introduction to the physical science of climate change, aimed towards economists. We begin by describing the physics that controls global climate, how scientists measure and model the climate system, and the magnitude of human-caused emissions of carbon dioxide. We then summarize many of the climatic changes of interest to economists that have been documented and that are projected in the future. We conclude by highlighting some key areas in which economists are in a unique position to help climate science advance. An important message from this final section, which we believe is deeply underappreciated among economists and thus highlight here, is that all climate change forecasts rely heavily and directly on economic forecasts for the world. On timescales of a half-century or longer, the largest source of uncertainty in climate science is not physics, but economics (Hawkins and Sutton 2009).

Basics of Climate Change Science

For most economic and scientific purposes, climate can be defined as the joint probability distribution describing the state of the atmosphere, ocean, and freshwater systems (including ice). Each of these systems is itself an extraordinarily high-dimensional system, so it is appealing to work with summary statistics such as *global mean surface temperature* or temperature distributions for major cities. Indeed, global mean surface temperature is intimately tied to the fundamental physics of planetary energy balance that explain global warming. However, consumers of

climate science should recognize that such simplifications, while sometimes useful, do not capture the entire picture.

The idea that human activity could alter the climate has a long history, going back almost two centuries (for an overview, see Weart 2018). However, it took focused research during the second half of the 20th century to achieve the level of confidence we now possess that human activity is altering the climate (Stocker et al. 2013; US Global Change Research Program 2017). This confidence comes from many lines of evidence based on observations at Earth's surface and throughout different layers of the atmosphere and oceans, geological reconstructions of historical climates, and two centuries of physical theory. The null hypothesis that humans have had no influence on global climate is now easily rejected given available data (for example, Hegerl et al. 2007).

Planetary Energy Balance and Greenhouse Gases

Sunlight continuously enters our planet's atmosphere from space. In order for the earth to maintain a stable surface temperature, this flow of incoming energy must be balanced by a flow of energy leaving the atmosphere. For the Earth, about 30 percent of incident sunlight is immediately reflected back out to space from the surface or from clouds. The remaining 70 percent is absorbed by the Earth's surface and atmosphere, and must be balanced by the planet's own emission of infrared radiation to space, which intensifies with higher temperatures. Without greenhouse gases, the equilibrium global mean surface temperature would be –18°C (about 0°F)[1], fully determined by the Sun's temperature, the Earth's distance from the Sun, and the reflectivity (also known as "albedo") of the Earth. If a larger flow of energy somehow were to reach the Earth's surface—for example, if the Sun were to grow in brightness, or the Earth to decline in albedo—the planet would heat up until the additional outgoing flow of infrared radiation exactly offset this new source of energy.

Greenhouse gases distort Earth's energy balance because they are transparent to incoming visible and ultraviolet sunlight but absorb infrared radiation, hindering the return flow of this energy from the surface and the lower atmosphere into space. When a greenhouse gas molecule intercepts infrared radiation headed from the surface to space, the absorbed energy is re-emitted in all directions, sending some energy that might otherwise have escaped to space back down to the surface of the Earth. This causes the surface and lower atmosphere to warm, increasing their emission of infrared radiation slightly. Equilibrium is re-established when the intensified outgoing infrared radiation is sufficient to offset the trapping effects of the greenhouse gases.

Because of the presence of greenhouse gases, the average height in the atmosphere from which infrared radiation can escape to space and contribute to balancing the planet's energy budget is not the Earth's surface; it is a level of the atmosphere known as the "effective radiating level." At present, Earth's effective

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[1] To convert any temperature *change* from Celsius to Fahrenheit, multiply by nine-fifths—so 2°C of warming is $2 \times 9/5 = 3.6°F$ of warming. To convert a temperature in *levels* from Celsius to Fahrenheit, multiply by nine-fifths and then add 32—thus a day with level temperature of $30°C = 30 \times 9/5 + 32 = 86°F$.

radiating level occurs at about 5.5 km altitude; on average, this level has the necessary temperature of about –18°C—the same that the Earth's surface would have in the absence of greenhouse gases. The relationship between temperature and altitude in Earth's lower atmosphere—on average about 6°C/km—makes the surface nearly 33°C (about 59°F) warmer than this level.

When greenhouse gases are added to the atmosphere, the first reaction is that the height of the effective radiating level moves upward. This temporarily leads to a decrease in the amount of radiation escaping from the Earth to space; but conservation of energy implies the surface and lower atmosphere must then warm up, so the higher (and originally cooler) effective radiating level would warm to the equilibrium temperature of –18°C. In the absence of additional feedbacks, doubling carbon dioxide concentrations would lead to the effective radiating level being about 200 meters higher, which in turn would lead to an equilibrium surface warming of about 1.2°C (Hansen et al. 1981).

However, the warming surface and atmosphere trigger feedbacks, which change the shift in effective radiating level and surface temperature associated with a given change in greenhouse gas concentration. Estimates of equilibrium climate sensitivity (the long-term, equilibrium response to a doubling of CO_2 concentrations) that include atmospheric and sea ice feedbacks are generally 2.0–4.5°C (3.6–8.1°F) (Collins, Knutti et al. 2013). The most important feedback involves water vapor: a warmer atmosphere is a more humid atmosphere, and water vapor is the most powerful natural absorber of longwave infrared radiation. Other important feedbacks involve sea ice (which reflects incoming solar energy), clouds (which can both trap heat and reflect incoming solar energy), and the response to warming of the ocean and land ecosystems (which drive most of the flow of CO_2 out of the atmosphere and can also affect albedo).

Because greenhouse gases alter the climate by changing the radiative properties of the atmosphere, their influence is measured in units of "radiative forcing," defined as the extent to which the human-generated stock of gas distorts the net flow of radiation into the atmosphere on average (incoming minus outgoing), relative to a preindustrial baseline. For example, a rise in atmospheric CO_2 concentrations from the historical baseline of 278 parts per million to the current (as of 2018) level of 409 parts per million exerts about 2.1 W/m^2 of radiative forcing. For reference, the energy from the sun reaching the top of Earth's atmosphere is 342 W/m^2, and central estimates of the equilibrium warming associated with a change in radiative forcing are about 0.8°C per W/m^2; thus the equilibrium warming associated with the current level of CO_2 forcing is about 1.6°C above the preindustrial baseline.

Radiative forcing by greenhouse gas emissions does not translate immediately into surface warming, in part because the deep ocean takes centuries to warm and, through exchange of heat with the surface ocean, slows overall warming. Nonetheless, modeling experiments indicate that—because of the relative timescales over which the planet warms and CO_2 is naturally removed from the atmosphere—most of the warming associated with a marginal emission of CO_2 occurs within a couple decades and persists for millennia (Joos et al. 2013). Thus, climatic changes experienced today are a result of both relatively recent emissions

and also cumulative emissions during the past centuries of fossil fuel combustion and past millennia of deforestation.

Establishing Baseline Climates

Within climate science, *paleoclimatology* is a well-developed subfield that focuses on the reconstruction of historical climates, thus setting a baseline for explaining climate changes. For examples, gases trapped in air bubbles of ice contain information on atmospheric chemistry at the moment they froze (for example, Luthi et al. 2008); the width of tree rings reflect growing-season temperatures and rainfall (for example, Jones et al. 2009); microscopic fossils in salt-marsh sediments reflect changes in salinity, and thus in local sea level (for example, Edwards and Horton 2000); and the relative abundance of different isotopes of oxygen in ocean sediments reflect the extent of "ice ages" because polar ice sheets lock up lighter isotopes, thereby restricting their supply to the deep ocean (for example, Cramer, Miller, Barrett, and Wright 2011). In some cases, physical data can be corroborated with observations in historical records, such as records of cyclone-caused shipwrecks maintained by insurers (Trouet, Harley, and Domínguez-Delmás 2016). While most proxies and historical observations are inherently local, spatio-temporal statistical methods and comparison to physical models can be used to estimate global mean values of quantities such as surface temperature and sea level from local data.

Figure 1 presents reconstructions of atmospheric CO_2 concentrations, global mean surface temperature, and global mean sea level over two different timescales. In the context of the last 11,600 years, known as the Holocene Epoch, the recent sharp jump in atmospheric CO_2 concentrations is quite striking and is unequivocally explained by human-caused emissions (Luthi et al. 2008; MacFarling Meure et al. 2006). The higher-resolution, post-1958 observational record from Mauna Loa in Hawaii also reflects higher-frequency patterns of largely natural variability, like the seasonal cycle and inter-annual variability in the strength of the land and ocean carbon sink (Keeling et al. 2001). The Holocene temperature record reveals a long-term decline, caused by slow variations in Earth's orbit, that began around 5,500 years ago (Marcott, Shakun, Clark, and Mix 2013).[2] The post-1850 reconstruction from direct observations reveals that this decline was interrupted in the 20th century, by a rise totaling about 1.0°C from the late-19th to the early-21st century (Rohde et al. 2013). This rise, which is superimposed by a spectrum of

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[2]The exact timing of the Holocene decline is currently contested. Marsicek, Shuman, Bartlein, Shafer, and Brewer (2018) suggest that the global analysis underlying Figure 1 is seasonally biased and conceals a more complex pattern, at least in North America and Europe, where their analysis suggests that summer temperatures declined starting around 5,500 years ago, but that winter temperatures did not cool until about 2,000 years ago. Some other, very long term cycles include Milanković cycles, which are global periodic climate changes driven by variations in the orientation of Earth's axis of rotation (19,000- and 23,000-year periods), the tilt of Earth's axis of rotation (41,000-year periods), and the shape of Earth's orbit around the Sun (roughly 100,000- and 400,000-year periods) (Berger 2012). Changes in incoming solar radiation caused by these cycles, amplified by natural feedbacks, serve as the pacemaker for ice ages over the last 2.6 million years.

Figure 1

Atmospheric CO$_2$ Concentrations, Global-Mean Surface Temperature, and Global-Mean Sea Level
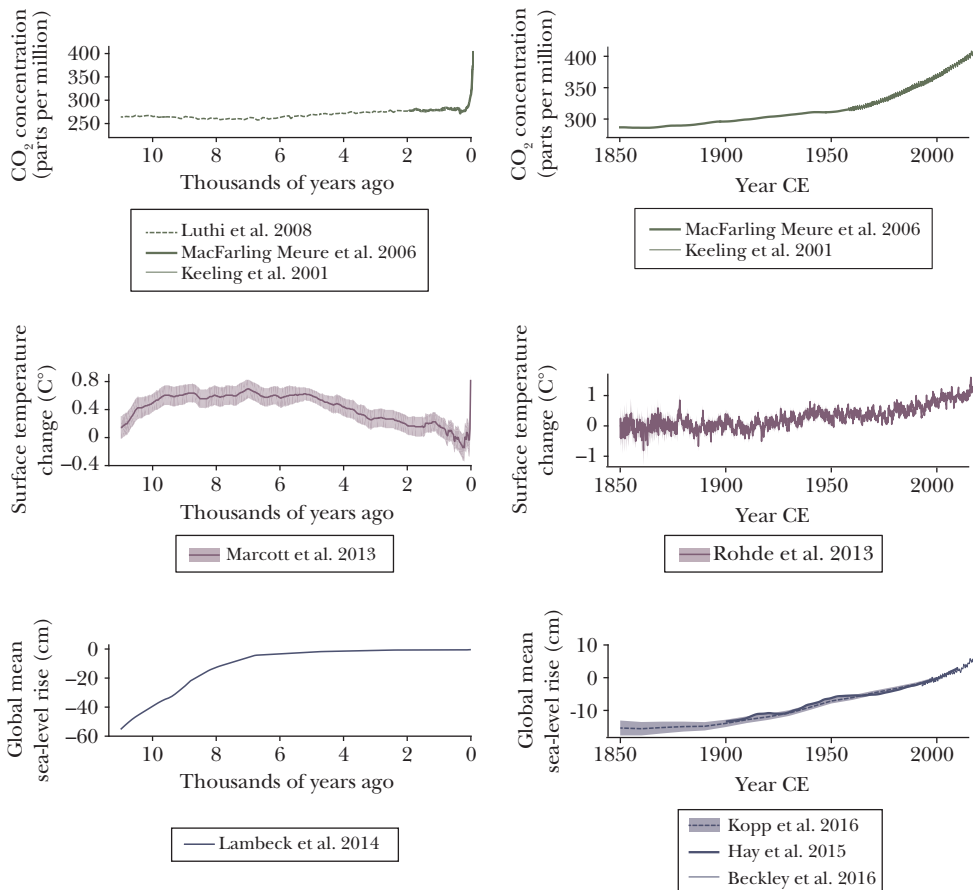
Data: Luthi et al. (2008); MacFarling Meure et al. (2006); Keeling et al. (2001); Marcott, Shakun, Clark, and Mix (2013); Rohde et al. (2013); Lambeck, Rouby, Purcell, Sun, and Sambridge (2014); Kopp, Kemp et al. (2016); Hay, Morrow, Kopp, and Mitrovica (2015); Beckley et al. (2016).
Note: The figure shows historical atmospheric CO$_2$ concentrations from ice cores and direct measurements (top), reconstructed historical global mean surface temperatures relative to the 1850–1900 average (middle), and reconstructed global mean sea level relative to the 1991–2009 average (bottom), over the last 11,000 years (left) and since 1850 CE (right). Shaded areas are 95 percent confidence intervals.

higher-frequency variability, some internal to the climate system and some driven by changes in forcing, is well-explained by the response to human-caused emissions.

Sea level responds more sluggishly than temperature to changes in forcing, because both the oceans (which expand when they absorb heat) and ice sheets (which can shrink in response to warming temperature) are large systems with an ability to absorb tremendous quantities of heat while warming only modestly. The first half of the Holocene is characterized by relatively rapid sea-level rise, which

ended with the final disappearance of Laurentide Ice Sheet in North America. This rise was a delayed response to about 5°C warming since the thermal nadir of the last ice age (about 21,000 years). The twentieth-century sea-level rise was the fastest in at least 2,800 years, and the last quarter-century was characterized by a rate about twice as fast as the 20th century average (Sweet, Horton, Kopp, LeGrande, and Romanou 2017).

In general, a core challenge to determining whether humans are changing the climate is assessing whether systematic changes in the behavior of the climate system are explained or confounded by the sources of natural variation. As one example of such natural variation, El Niño–Southern Oscillation is the dominant pattern in the global climate at annual frequencies and has occasionally been studied by economists (for example, Hsiang and Meng 2015). Other longer-term ocean-related oscillations include the North Atlantic Oscillation, which varies on seasonal, annual, decadal, and centennial timescales (Hurrell 1995; Trouet, Scourse, and Raible 2012); the Pacific Decadal Oscillation (Mantua and Hare 2002); and the Atlantic Multidecadal Oscillation (Clement et al. 2015). Climate models form the basis for inference in this setting, seeking to separate a secular trend signal from these oscillating sources of noise.

Climate Models

Climate models mathematically represent physical understanding of the climate system. They fall along a hierarchy of complexity from simple models that capture key aspects of the longer-term, global-scale response to detailed, full-complexity Earth system models that provide greater insight into processes at finer temporal and spatial scales (Hayhoe et al. 2017).

The simplest climate change models, called energy balance models, can simulate millennia of global mean climate change in a single second on a laptop. These models are based on a budgeting of sunlight and thermal energy in the atmosphere, as well as the role of key feedbacks. Early pen-and-paper versions of such models date back to the work of Svante Arrhenius in the 1890s; by the 1960s, they had also been adapted to include a single spatial dimension representing the vertical structure of the atmosphere, which allowed the models to describe vertical motions of air (Manabe and Strickler 1964).

In the 1960s, the equations of fluid dynamics were incorporated to produce early atmospheric "general circulation models" that capture both the three-dimensional structure and dynamical evolution of the global atmosphere (for example, Manabe and Smagorinsky 1965). Later generations of models elaborated their representation of the ocean, as well as of sea ice, land surfaces, and atmospheric chemistry. These general circulation models[3] were the ancestors of today's full-complexity Earth system models, which also endogenize vegetation dynamics and the carbon cycle. Full-complexity Earth system models represent the best tools available for simulating spatial patterns of the climate response, but they have several drawbacks.

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[3] As general circulation models evolved to include more than just the fluid dynamics of the ocean and the atmosphere, the acronym GCM was sometimes adapted to stand for "global climate model."

First, they are computationally expensive, taking several hours on a high-performance computing cluster just to simulate one year of climate. Second, although such models provide fairly high spatial resolution—with grid cells that are roughly 100 km along a side in the generation of models used in the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC)—this resolution may still be inadequate for capturing details relevant to many economic impacts. Third, detailed models may produce baseline climate projections that differ from observed historical patterns. To address these last two issues, the climate science community has developed post-processing techniques for bias-correction and spatial "downscaling," thus increasing the spatial resolution of the final output. Such techniques include both statistical approaches (like using quantile regressions to mimic historical variability around the mean) and physical modeling approaches that embed higher-resolution regional climate models within boundary conditions set by a global model (as in Wood, Leung, Sridhar, and Lettenmaier 2004). In addition, cutting-edge climate models are run at increasingly higher resolutions; some of the most recent models have resolutions below 50 km × 50 km, and in some cases can achieve local resolutions as high as 10 km × 10 km.

Within the context of a single climate model simulation, uncertainty arises from the imperfect representation of physical processes—that is, from structural and parametric uncertainty—as well as from the imperfectly known initial conditions of a model run. As famously discovered by Lorenz (1963) in early numerical weather models, tiny errors in initial conditions can produce dramatically different forecasts within the same model, chaotic behavior known colloquially as "the butterfly effect." This endogenous chaotic behavior turns out to be more difficult to predict than global average conditions, which are tightly constrained by energy budgets. As a result, climate modeling teams usually run their model multiple times with perturbed initial conditions, creating a collection of results known as an *initial-conditions ensemble*. Individual realizations of the model are never interpreted as literal forecasts; rather, the ensemble as a whole is thought to capture statistical properties of the climate system. Indeed, most climate scientists generally avoid the terms "forecast" and "prediction," preferring instead the term "projection" to describe a simulation of future climate under an assumed emission scenario. Producing decadal projections with global climate models is a frontier research area, with one of the key challenges being aligning the internal variability of a climate model with the internal variability of the real climate (Meehl et al. 2009).

Emissions of Radiative Pollutants

Not all the greenhouse gases emitted by humans remain in the atmosphere today; a substantial fraction has been absorbed by carbon sinks on land (like plants and soil) and in the ocean (for example, by phytoplankton and chemical dissolution). If all 1434 Gt of fossil CO_2 emitted since 1750 had stayed in the atmosphere, the current atmospheric CO_2 concentration would be about 475 ppm rather than the observed 409 ppm, even without considering emissions from deforestation.[4]

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[4]One part-per-million CO_2 in the atmosphere is equal to about 7.8 Gt CO_2 in physical mass.

Table 1

Historical and Top 15 Current Emissions of Carbon Dioxide from Fossil Fuel Combustion and Cement Production

Country	Cumulative 1751–2014 (gigatonnes CO_2)	% of Global	Emissions 2014 (gigatonnes CO_2)	% of Global	Emissions per capita (tonnes CO_2), 2014
China	174.7	12%	10.3	30%	7.5
United States	375.9	26%	5.3	15%	16.2
India	41.7	3%	2.2	7%	1.7
Russia / USSR	151.3	11%	1.7	5%	11.9
Japan	53.5	4%	1.2	4%	9.6
Germany	86.5	6%	0.7	2%	8.9
Iran	14.8	1%	0.6	2%	8.3
Saudi Arabia	12.0	1%	0.6	2%	19.5
South Korea	14.0	1%	0.6	2%	11.7
Canada	29.5	2%	0.5	2%	15.1
Brazil	12.9	1%	0.5	2%	2.6
South Africa	18.4	1%	0.5	1%	9.1
Mexico	17.5	1%	0.5	1%	3.8
Indonesia	11.0	1%	0.5	1%	1.8
United Kingdom	75.2	5%	0.4	1%	6.5
World	**1,434.0**	**100%**	**34.1**	**100%**	**4.7**

Source: Boden, Marland, and Andres (2017).

However, cumulative emissions of CO_2 are nonetheless a useful metric, as the CO_2-caused warming is approximately proportional to cumulative emissions (Allen et al. 2009), with every trillion tons of CO_2 causing about 0.2–0.7°C of warming.

Table 1 presents the estimated cumulative emissions of CO_2 from fossil fuels and cement production during 1751–2014, as well as the flow of emissions in 2014 (Boden, Marland, and Andres 2017). The United States is responsible for over one-fourth of historical emissions, followed by China (12 percent) and Russia (11 percent, including the former Soviet Union); together with Germany (6 percent) and the United Kingdom (5 percent), these five countries account for 60 percent of historical emissions. However, if one examines flows today rather than the stock of historical emissions, the picture is changing; China (30 percent) dominated emissions in 2014, followed by the United States (15 percent), India (7 percent), Russia (5 percent), and Japan (4 percent). Germany is the largest emitter in the European Union (2.1 percent), with the EU28 collectively ranking third in global CO_2 emissions, responsible for about 10 percent (Janssens-Maenhout et al. 2017). High national emissions reflect high carbon intensity per capita (per-capita emissions are 16.2 tonnes/year in the United States, 3.4 times the global average), high population levels (per capita emissions in India, the third-leading emitter, are about one-third the global average), or a mix of both factors (per-capita emissions in China are about 60 percent more than the global average).

These metrics do not include CO_2 emissions from deforestation, which are significant: Pongratz and Caldeira (2012) estimate that these accounted for about

230 gigatonnes of CO_2 from 800–1850, and 425 gigatonnes of CO_2 from 800–2006, compared to about 1,175 gigatonnes of CO_2 from fossil fuels over this latter time period (Boden, Marland, and Andres 2017). At present, the ratio of fossil fuel to land use emissions is about 7.6 (Le Quéré et al. 2018).

These metrics also do not include emissions of non-CO_2 greenhouse gases and other climate-altering pollutants. The climatic impact of an emission depends on both its radiative forcing of the molecules emitted and their lifetime in the atmosphere. For example, methane survives for only 12 years on average in the atmosphere before breaking down into CO_2 and water, whereas a substantial fraction of emitted CO_2 lasts for millennia. Thus, while methane has large radiative impact per molecule per year, the integrated lifetime impact of a marginal molecule of methane emissions is partially offset by its short lifetime.[5] Blanco et al. (2014) provide a discussion of non-CO_2 emissions.

Emissions of particulate matter and aerosol precursors (like sulfur dioxide) also influence the radiative balance of the atmosphere. Both pollutants lead to the formation of aerosols—particles that are solid or liquid, not gases, but which are small enough to remain aloft in the atmosphere for substantial periods of time (days in the lower atmosphere; years in the stratosphere). Most aerosols reflect incoming sunlight, leading to surface cooling (negative radiative forcing), but some, notably black carbon, absorb solar energy and increase warming. Through their effects on cloud physics, aerosol emissions have complex regional consequences for precipitation that are distinct from the effects of greenhouse gases (Rosenfeld et al. 2008). Because the spatial distribution and net radiative effects of aerosols are difficult to monitor, and change more quickly than gases, the overall radiative impact of aerosols is highly uncertain and remains an important open question in climate change science. The global average effective radiative forcing of aerosols is estimated to be between –1.9 and –0.1 W/m^2—opposite in sign and between about 5 percent and 90 percent of the forcing from CO_2 (Boucher et al. 2013).

As one more level of complexity, coal combustion emits both CO_2 and aerosol pollution, which leads to a tradeoff of timescales: burning less coal reduces particulate matter and sulfur dioxide emissions which is directly beneficial to human health, but also leads to a short-term increase in warming due to the reduction in aerosol emissions, even though the long-term effect of reduced CO_2 emissions is a substantial reduction in warming (Wigley 2011). Similarly, efforts to target reductions in particulate pollution from coal power plants without tackling CO_2 emissions will lead to climate warming (Westervelt et al. 2015).

Emissions Scenarios

There are many climate modeling research programs, each of which develop, maintain, and run global climate models whose outputs are compared against one

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[5]Methane, with an atmospheric concentration of about 1.8 ppm, currently exerts a forcing of about 0.5 W/m^2; nitrous oxide, at 0.3 ppm, exerts a forcing of about 0.2 W/m^2, and fluorinated gases like chlorofluorocarbons and hydrofluorocarbons, with concentrations less than 1 part per billion, exert forcing of about 0.3 W/m^2.

another. The Coupled Model Intercomparison Project (CMIP) (Taylor, Stouffer, and Meehl 2012) is the largest comparative effort, and plays a major role in informing the assessment reports of the Intergovernmental Panel on Climate Change. To ensure that model outputs are comparable across groups, standardized emissions scenarios are used as inputs to all models. The latest effort, CMIP Phase 5 (CMIP5), used a range of emission scenarios, known as the Representative Concentration Pathways (RCPs), that exogenously prescribe the flow of human-caused emissions over the coming decades. These emissions scenarios, which begin in 2005, are labeled by the overall radiative forcing (in W/m^2) that occurs in 2100 in each scenario. RCP 8.5 has the strongest forcing, with CO_2 emissions nearly doubling from their current levels by 2050 and continuing to rise thereafter; RCP 4.5 has a moderate forcing, with CO_2 emissions stabilizing at close to their current levels through the middle of the century and declining thereafter, reaching about 40% of their current levels by 2080; and RCP 2.6 has the weakest forcing, with CO_2 emissions declining immediately, to less than a third of the current levels by 2050, and becoming net-negative during the 2080s. In RCP 8.5, atmospheric CO_2 concentration climbs to 541 ppm by 2050 and 936 ppm by 2100; in RCP 4.5, to 487 ppm by 2050 and 538 ppm by 2100; and in RCP 2.6, to 443 by 2050, declining to 421 ppm by 2100. Below, when we discuss "high-", "moderate-" and "low-" emissions scenarios, we are referring to RCP 8.5, 4.5, and 2.6, respectively.

Observed and Projected Climate Changes in the Modern World

In this section, we describe how historical changes in the climate are identified and attributed to human activity, as well as climate changes that are projected to occur. Interested readers should consult the IPCC Fifth Assessment Report (Stocker et al. 2013), USGCRP (2017), and the readings cited below for additional details.

Detection and Attribution of Climate Change

Over the last several decades, a core objective of climate science has been to detect changes in the climate and to determine whether these changes can be attributed to human activity. *Detection* refers to the empirical problem of determining whether there has been an actual shift in the joint distribution of environmental variables that we refer to as the climate. *Attribution* refers to the inferential problem of assigning a cause to the observed changes (Bindoff et al. 2013). Attribution studies generally simulate what counterfactual climates would look like in the absence of human activity, altering the model parameters that describe human inputs to the climate. Thus, for example, human emissions of greenhouse gases and aerosols might be eliminated in a model's "control" simulation. If it is not possible, or sufficiently unlikely, that these human-free simulations can account for observed changes in the climate, then scientists attribute these changes to human activity.

The scientific community is in broad and strong agreement that overall, human activity has already substantially altered the global climate and that continued changes should be expected as emissions of greenhouse gases continue (Stocker et al. 2013; USGCRP 2017). The vast majority of actively publishing researchers now acknowledge

Table 2

Statements of the Intergovernmental Panel on Climate Change (IPCC) on Detection and Attribution of Global Climate Change

First Assessment Report (1990)	"Unequivocal detection of the enhanced greenhouse effect from observations is not likely for a decade or more."
Second Assessment Report (1995)	"The balance of evidence suggests a discernible human influence on global climate."
Third Assessment Report (2001)	"Most of the observed warming over the last 50 years is *likely** to have been due to the increase in greenhouse gas concentration."
Fourth Assessment Report (2007)	"Most of the observed increase in global average temperatures since the mid-20th century is *very likely* due to the observed increase in anthropogenic greenhouse gas concentrations."
Fifth Assessment Report (2013)	"It is *extremely likely* that human influence has been the dominant cause of the observed warming since the mid-20th century."

Source: The IPCC Assessment Reports can be found at https://www.ipcc.ch/publications_and_data/ publications_and_data_reports.shtml.
* The uncertainty language used by the IPCC is precisely defined: *likely* refers to an assessed probability of at least 66 percent, *very likely* implies at least 90 percent, and *extremely likely* means at least 95 percent.

the strength of the evidence implicating a human-caused signal in climate change (Cook et al. 2016). The agreement in the scientific community has grown stronger over the last quarter century, reflected in the IPCC's increasingly strong statements regarding the detection and attribution of global warming shown in Table 2. Some of the public confusion regarding the strength of scientific evidence appears to have been sown intentionally. For example, a textual analysis of ExxonMobil documents from 1977–2014 indicates that internal documents generally acknowledged that climate change is real and human caused while public-facing documents did not (Supran and Oreskes 2017).

Figure 2 (which is best viewed in the color version of this article available at the JEP website) shows some of the most important evidence in support of the conclusion that human emissions are causing global temperatures to rise. In the left panel, red [or upper light grey] bands indicate the range of global mean surface temperature simulated in 90 percent of climate models that exogenously impose observed human emissions. Blue [or lower light grey] bands indicate the analogous range for the same models but in a "control" simulation that imposes only natural forces. Observed temperatures, indicated by the black line, began to separate from the envelope of control simulations in the 1980s and now lie far outside this range. In contrast, observed temperatures are fully consistent with the range of temperatures simulated when human emissions are included. We note that this consistency extends not just to global mean surface temperature, but also to changes in stratospheric temperature and ocean heat content. Thus, it is extremely difficult to explain current temperatures in the absence of human activity. The

Figure 2

Average Annual Global Mean Surface Temperature, Compared to Distributions of Climate Model Simulations
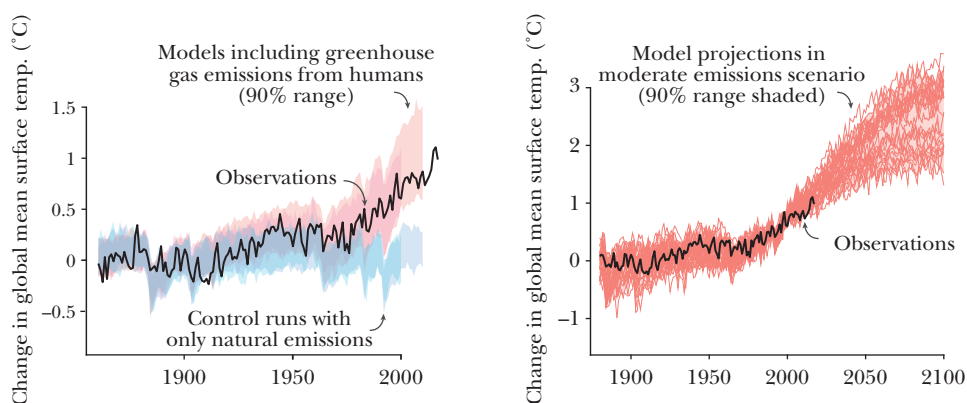
Sources: Data comes from Jones, Stott, and Christidis (2013), Morice, Kennedy, Rayner, and Jones (2012), and Taylor, Stouffer, and Meehl (2012).

Note: This graph is best viewed in color; the electronic version of this article available at the JEP website is in color. The heavy black line shows observed average annual global mean surface temperature. The red [or light grey] distributions are exogenously "treated" with anthropogenic greenhouse gas emissions, while the blue [or light grey] distributions (shown only in the left panel) are "control" runs that only contain natural forcings. In the left panel, climate model distributions are from the Third Coupled Model Intercomparison Project (CMIP3) published in 2007 and displayed until 2000, and CMIP5 published in 2013 and displayed until 2010. In the right panel, all climate model projections come from CMIP5 in the moderate emissions scenario (RCP 4.5). Temperatures shown are relative to the 1880–1900 average.

gradually increasing confidence of the scientific community can be understood by noting the envelope of model results published in association with the 2007 IPCC report (displayed ending in simulation year 2000) were less cleanly separated than those published in association with the 2013 IPCC report (displayed ending in 2010), although the separation visible through 2000 was already reflected in the IPCC's 2007 statement that temperatures were "*very likely* due to anthropogenic greenhouse gas concentrations" (Table 2).

It is now virtually certain (at least 99 percent probability) that the observed modern warming trend exceeds the bounds of natural variability (Bindoff et al. 2013). Furthermore, humans are *likely* (with at least 66 percent probability) responsible for 0.6°C–0.8°C of the observed 0.6°C of warming over 1951–2010. Values greater than 0.6°C are possible for the anthropogenic contribution because of the possibility that natural forcing and variability could otherwise impose a slightly negative baseline trend (for example, as a result of volcanic eruptions), a pattern which is visible in the control runs of Figure 2.

Temperature Changes

Since the late 19th century, global mean surface temperature has increased by about 1.0°C, with the trend accelerating after 1980. Almost every location on the planet has exhibited an upward temperature trend over this period (Wuebbles et al. 2017). Warming has also been substantially faster over land than the ocean—between 1880–1900 and 1997–2017, the land has warmed 1.4°C (2.5°F) on average while the oceans warmed roughly 0.6°C (1.1°F) (GISTEMP Team 2018).

As one would expect, given the array of factors affecting temperatures, the overall rise in temperatures has not been smooth over time or homogenous across space. For example, warming was dampened in the 1950s–1970s, most likely as a result of both aerosol emissions, which reflected sunlight away from the planet (Maher, Gupta, and England 2014), and natural variability. Since 1980, the most rapid warming has occurred in the far north, where the replacement of highly reflective summer sea ice with dark, open ocean rapidly increases the absorption of sunlight and local warming (Serreze and Barry 2011).

A heavily discussed period of slowed average warming over 1998–2013, the so-called "hiatus," now appears fully consistent with the natural variance of the climate system (Cahill, Rahmstorf, and Parnell 2015), as can be readily seen in the overlay of simulated and observed temperature time series in the right panel of Figure 2. Relative to the distribution of simulations for the first decade of the 21st century, the observed values fall toward the low end of projections but never leave the envelope of expected variations. However, in addition to natural variability, it is thought that some model simulations warmed too quickly because the emissions scenarios in the RCPs underrepresented volcanic and human aerosol emissions after 2005 (Medhaug, Stolpe, Fischer, and Knutti 2017).

Based on the assessment of the Intergovernmental Panel on Climate Change of CMIP5 simulations, projected global mean surface temperature is likely to rise 0.9–2.3°C (1.6–4.1°F) above preindustrial levels (defined as the 1850–1900 average) by 2080–2100 under a low-emissions scenario, 1.7–3.3°C (3.1–5.9°F) under a moderate-emissions scenario (shown in Figure 2), and 3.2–5.4°C (5.8–9.7°F) under a high-emissions scenario (Collins et al. 2013).

As in the past, warming will be more rapid over land, where most economic activity occurs, compared to over the ocean. The only location on the surface that is projected by some models to cool is a very small portion of the North Atlantic Ocean just south of Greenland, where changing ocean circulation may induce cooling. Although warming will continue to occur fastest over the Arctic, average summer temperature will diverge from the historical range soonest in low-latitude regions, which experience lower historical variance. Figure 3A illustrates regional heterogeneities in the rate of warming (in °C) that are otherwise masked by globally averaged summary statistics. The map depicts the average warming at each location associated with a 1°C increase in global mean temperature; values greater than 1°C indicate rates of warming faster than the global mean, while values below 1°C indicate warming that is slower than the global mean.
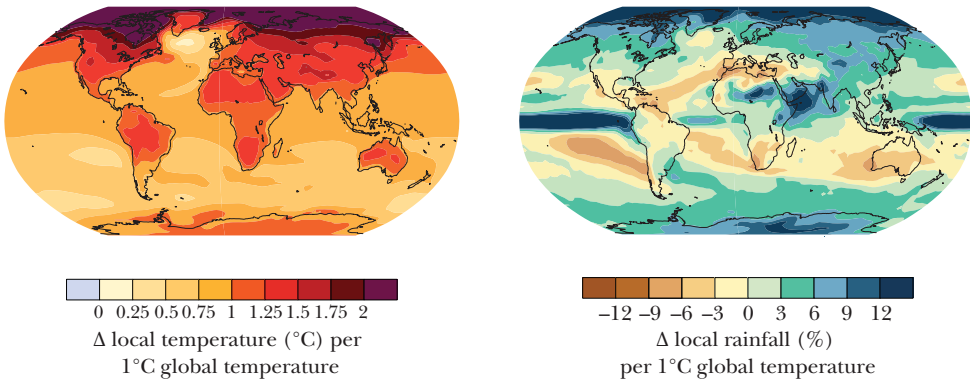
To help grasp the potentially transformative scale of these thermal changes, Figure 4, Panel A, plots the average summertime temperatures for the lower 48 US

Figure 3

Projected Change in Local Average Temperatures and Local Average Rainfall per 1°C of Warming in Global Mean Temperatures

A: Temperature change

B: Rainfall change

Source: Collins, Knutti, et al. (2013).
Note: Changes are differences in means between 1986–2005 and 2081–2100 in CMIP5 simulations of RCP 4.5, scaled by the overall change in global mean temperature. These heatmaps should be viewed in color. See the electronic versions on the JEP website.

states, adapted from Houser et al. (2015). The cluster in the bottom left of the figure in blue text indicates historically observed temperatures, while the cluster in the upper right of the figure in red text indicates average projected mean temperatures for 2080–2099 across models simulating a high-emissions scenario. This layout allows for projected future temperatures to be matched to historical analogs. For example, future summers in Vermont will be similar to historical summers in Maryland, summers in Connecticut will be similar to past summers in Arizona, future summer in New Jersey will be slightly hotter than historical Louisiana summers, and future summers in Georgia and Florida will be much hotter than anything previously experienced in the United States. As shown in Panel B, a similar analysis at the level of countries shows that future temperatures in Norway are projected to be similar to historical temperatures in Germany, future Mexico will be slightly hotter than historical Iraq, future Indonesia will be similar to historical Mali, and India and Thailand are projected to be hotter than any country presently on Earth.[6]

Precipitation Changes

A warmer atmosphere is capable of holding more water vapor, leading to an increase in overall average precipitation (rainfall and snowfall). Observed precipitation in the mid-latitude Northern Hemisphere has increased since the 1950s. Heavy precipitation events in particular have increased, most clearly in North America

[6] The Appendix available with this paper at http://e-jep.org shows current and projected annual average temperature for 166 countries.

Figure 4

Average Temperatures for Lower 48 US States Observed during 1981–2010 and Projected for 2080–2099 in a High Emission (RCP 8.5) Scenario.

A: States (USA)

2080–2099 high emission (RCP 8.5) scenario

WY MI WI   UT   SD CA   DC SC FL OK
OR CO RI   PA WV IA   KY   AL MS TX
ME WA NY MA   OH VA   IL DE MO   GA AR LA
VT ID MT ND CT   NM HI IN   MD NC
NH   MN   NV   NE   TN   KS AZ
NJ

1981–2010 (Historical)

WA   WI   PA   NM   MD   KS   MS
ME ID   MN   MA NV WV   NJ VA   MO HI   AZ OK   LA
OR   MT CO   CT   OH IN IL KY TN DC   GA   TX
VT   NH NY   RI UT   IA   CA   NC   AR   FL
WY ND   SD   NE   DE   SC
MI   AL

65   75   85   °F
    20   30   °C

June–July–August average temperature

B: Countries

2080–2099 high emission (RCP 8.5) scenario

USA   Brazil Haiti
Norway   S. Korea   Iran   Iraq Thailand   Mali
Mongolia   Turkey Australia Mexico   Bangladesh
Canada   China S. Africa   Uganda   India
Greenland   Russia UK France Japan   Saudi Arabia
Germany   Argentina Kenya   Indonesia
Spain   Honduras Nigeria Sudan

1981–2010 (Historical)

China   Honduras Sudan
Germany   S. Africa   Iraq Bangladesh
Norway   Japan   Argentina   Indonesia
Greenland   S. Korea   Uganda Haiti   Mali
Canada   UK   USA Iran Mexico   Saudi Arabia
Mongolia   Russia France   Australia Kenya   Nigeria
Turkey   Spain   Brazil   India
Thailand

25   45   65   85   °F
    0   10   20   30   °C

Annual average temperature

Note: Panel A displays summertime area-average temperatures adapted from Houser et al. (2015). Panel B displays population-weighted annual average temperatures, using data from Burke, Hsiang, and Miguel (2015). Markers are vertically jittered for readability.

and Europe, where the most data is collected (Hartmann et al. 2013). Both of these changes are consistent with those expected on a warming planet. However, because the atmospheric dynamics that govern precipitation involve both large motions of air masses and processes that occur at scales below the spatial resolution of many climate models, precipitation changes are considerably more challenging to model numerically than temperature changes. This difficulty, combined with the array of changes in temperature, wind, humidity and other factors that all affect when and where precipitation falls, have rendered projections of precipitation changes more complex and more uncertain than projections of temperature.

There is large heterogeneity in the sign of projected precipitation change, with many locations getting wetter while many others get drier. Precipitation dynamics are also strongly affected by internal variability—such as the El Niño–Southern

Oscillation—and projections for specific locations depend upon changes in large-scale patterns of atmospheric circulation (Collins et al. 2013). Figure 3B illustrates average changes in local rainfall for each 1°C increase in global mean temperature. At many locations, there is a large range of uncertainty for projected changes, with plausible projections allowing for no change. Simple summary statements like "dry regions are likely become generally drier and wet regions are likely to become generally wetter" hold well over the ocean, but are coarse descriptions of the complex precipitation changes that may occur over land (Greve et al. 2014). In the United States, the most robust projections are for a springtime drying of the Southwest, summertime drying of the Northwest, and increase in winter and spring precipitation in the Northeast, upper Midwest, and northern Great Plains (Houser et al. 2015).

Humidity Changes

Specific humidity is the total moisture content of air. *Relative humidity* is the ratio of specific humidity to a theoretical maximum moisture capacity, which rises exponentially as temperature increases. Since the 1970s, global mean specific humidity has increased; however, there is little evidence of an increase in global mean relative humidity and some evidence for a decline, possibly reflecting faster warming of the land than of the oceans, which are the primary source of atmospheric moisture (Sherwood and Fu 2014). Models that project that the largest increases in temperature on land also tend to predict the largest decreases in relative humidity (Fischer and Knutti 2013).

One reason that humidity is thought to be economically important is that it affects human health, since higher humidity levels make it more difficult for the human body to cool itself in hot conditions through sweating. One physical metric closely related to the combined effect of heat and humidity is *wet-bulb temperature*. Wet-bulb temperatures are measured using a ventilated thermometer wrapped in a wet cloth, and are strongly related to the experienced conditions of a sweating person.

Dangerously hot and humid conditions are projected to become dramatically more likely in several regions around the world (Sherwood and Huber 2010). For example, Houser et al. (2015) defined "dangerously hot and humid days" as those characterized by peak wet-bulb temperatures over 80°F. By this definition, dangerously hot and humid days are "typical of the most humid parts of Texas and Louisiana in the hottest summer month, and the most humid summer days in Washington and Chicago." Their analysis found that, in the southeastern United States, the population-weighted frequency of dangerously hot and humid days are projected to rise from 8 per year on average in 1981–2010 to 17–28 days per year over 2040–2059 in a moderate emissions scenario and to 40–70 days/year on average over 2080–2099 in a high emissions scenario (Houser et al. 2015).

Tropical Cyclones

Tropical cyclones are the class of phenomena that includes tropical storms, typhoons, hurricanes, and cyclones; these categories are distinguished by wind speed and the ocean basin where the storm occurs. Tropical cyclones are driven by the temperature difference between the warm ocean surface and cooler temperatures higher in the atmosphere. The warm ocean moistens overlying air, which rises and

cools, releasing energy and rain. Thus, climate change is thought to have countervailing effects on storms: warming sea surface temperatures fuel storms, but warming temperatures higher in the atmosphere may suppress them (Knutson et al. 2010).

Tropical cyclone formation and storm trajectory depend on myriad additional factors, especially wind patterns, which introduce additional complexity into projections of their future changes. Furthermore, inconsistent historical data on storms in the open ocean prior to the satellite era make inferences difficult. However, there is evidence that the frequency and intensity of the strongest storms in the Atlantic have been increasing since the 1970s, and some evidence that humans contributed to this change (Walsh et al. 2016).

Efforts to model all of these factors together broadly agree that the frequency of intense tropical cyclones (such as category 4 and 5 hurricanes), as well as the average intensity of their associated rainfall, is projected to increase with warming (Kossin et al. 2017). The effect on total number of storms remains less certain, though most studies suggest a stable or decreasing quantity of lower-intensity storms. The effect of climate change on *storm tracks* (the paths that storms take toward land) is uncertain and may offset or enhance the effect of increased storm intensity in some regions.

The spatial distribution of these changing risks is heterogeneous. For example, systematic changes in the spatial distribution of storm tracks within an ocean basin may reallocate cyclone risk between populations, even if the overall frequency of storms does not change. Across ocean basins, models generally agree in projecting substantial increases of storm intensity in the West Pacific, affecting East Asia and Oceania, with some decreases in activity occurring in the Indian Ocean, affecting Southern Asia and East Africa. Projecting changes in the North Atlantic, which affects Central and North America, has been more challenging; the greatest scientific uncertainty persists for this area (Knutson et al. 2010; Christensen et al. 2013).

Humans may also affect the genesis and growth of tropical cyclones through the regional effects of aerosol pollution. Aerosols aloft may cool local sea surface temperatures, by reflecting sunlight before it reaches the sea surface, as well as heat higher levels in the atmosphere, by absorbing sunlight when the particles are dark-colored. Both of these effects generally work to weaken storms, and storm activity in recent decades may have been greater had greenhouse gas co-pollutants been absent (Walsh et al. 2016).

Sea-Level Rise

Global mean sea-level rise is driven by two processes: an increase in the volume of the water already in the ocean, which occurs as the water warms and expands, and an increase in the mass of water in the ocean, primarily from the melting of ice on land. Since 1900, global mean sea level has increased by about 18–21 cm, with the rate of rise since about 1990 being 2–2.5 times faster than during the preceding nine decades. A substantial fraction of this rise is attributable to human-caused climate change (Sweet et al. 2017). Regional sea-level changes can differ substantively from this global trend, modulated by changes in currents and winds; changes in Earth's gravitational field, rotation, crust, and mantle that occur as land ice changes; and changes in the height of land that result from compaction of sediments, plate

tectonics, the ongoing mantle response to historical land ice changes, and other factors (Kopp, Hay, Little, and Mitrovica 2015). Historic sea-level rise has led to a detectable increase in the frequency of coastal flooding, in some cases by more than an order of magnitude (Sweet and Park 2014).

Due to the slow response time of the oceans and ice sheets, sea-level rise is fairly insensitive to alternative emissions scenarios for the first half of this century. Across studies, median projections of future global mean sea-level rise are 20–30 cm during 2000–2050, with less than a 5 percent chance of exceeding 50 cm (for a review see Horton et al. 2018).

After 2050, projections become more deeply uncertain, due to both uncertain human emissions and the uncertain response of the polar ice sheets (Kopp, DeConto, et al. 2017). Median projections for 2000–2100 range from 40–80 cm for a low-emissions scenario to 70–150 cm for a high-emissions scenario. However, global mean sea-level rise of as much as 250 cm by 2100 cannot be ruled out. For reference, the last time global temperatures were about as high as they currently are (about 125,000 years ago), global mean sea-level was about 6–9 meters higher than today. Some coastal areas will be inundated permanently; others will be protected by additional investments to be incurred. The resulting increase in the frequency of tide- and wave-driven flooding is expected to render some low-lying island states uninhabitable (Storlazzi et al. 2018).

In addition to increasing average high-tide levels, a major economic consequence of sea-level rise results from its interaction with tropical cyclones and extratropical cyclones (like the "nor'easter" storms that sometimes hit the northeastern United States). Storm surges that occur during these storms can impose major costs and sea-level rise adds roughly linearly to peak storm surge height. Figure 5 illustrates the joint effects of projected sea-level rise and changing tropical cyclone activity on the flood risk of Miami, FL, and New York, NY (Hsiang et al. 2017). The extent of areas expected to flood with a 1 percent annual probability increases substantially after 2050 for Miami but much earlier in the century for many regions of New York.

Droughts and Floods

By altering temperature and precipitation patterns, climate change alters the frequency and intensity of extreme moisture conditions, such as droughts and floods. There is a limited but increasing ability to attribute intensifying extreme floods and drought to human activity. For example, Emanuel (2017) estimated that climate forcing by humans amplified the probability of rainfall experienced by Texans during Hurricane Harvey six-fold.

Since 1950, the likelihood of drought has increased in the Mediterranean and West Africa and decreased in central North America and northwestern Australia (Hartmann et al. 2013); again, assessment of the human role in this trend is challenging (Bindoff et al. 2013). Some measures of drought in the United States have increased due to warming, which increases evaporation and exhibits an anthropogenic signal (Wehner, Arnold, Knutson, Kunkel, and Legrande 2017). In model projections, the frequency of droughts tends to increase in dry regions (Collins et al. 2013), which are also projected to expand. Prolonged hot and dry periods are

Figure 5

Areas Projected to Experience Floods at Least Once every 100 Years on Average (1% annual risk) in Miami, FL, and New York, NY
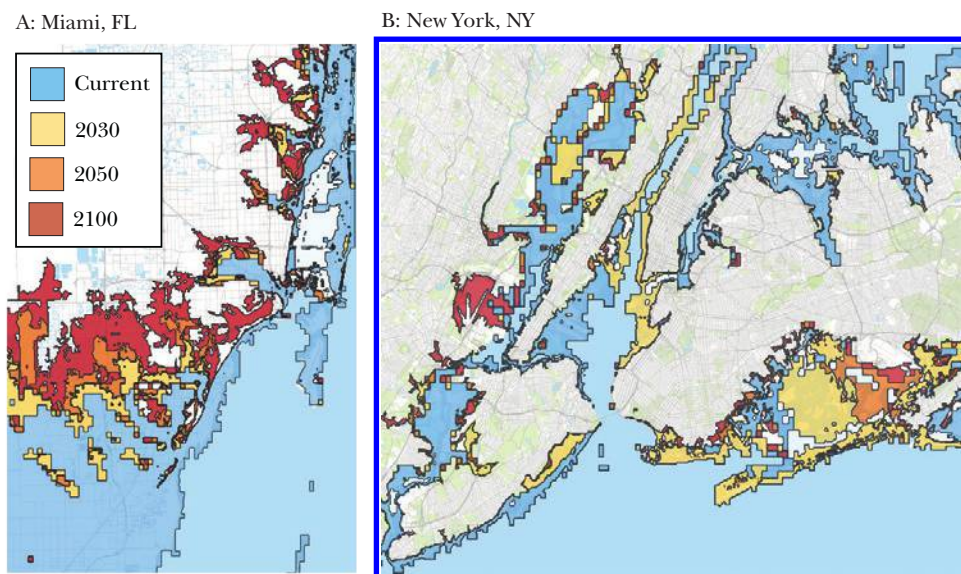
Source: Hsiang, Kopp, Jina, Rising, et al. (2017).
Note: These projections account for median projected sea-level rise and for projected changes in tropical cyclone intensity in a high-emission (RCP 8.5) scenario.

projected to become substantially more frequent in many grassland areas with low agricultural productivity, regions that today often depend on livestock production (Bell, Sum, Longmate, Tseng, and Hsiang 2018). In regions with more vegetation, amplifying oscillations between heavy rainfall and drought is projected to increase the frequency of wildfires (Abatzoglou and Williams 2016) because vegetative fuel grows rapidly during wet periods then becomes flammable during dry periods.

The observed increase in heavy precipitation suggests that climate change is contributing to increasing rain-driven flood damages (Jiménez Cisneros, Oki, et al. 2014). Projected increases in heavy precipitation and shorter-lived snowpack are likely to further increase the frequency of inland flooding. In addition to changing rainfall patterns, rising sea levels amplify the frequency of coastal flooding (Buchanan, Oppenheimer, and Kopp 2017). Changing patterns of flood risk are of particular economic importance, as flooding is recognized as one of the costliest classes of disaster globally (Swiss Re Institute 2018). However, in many cases, the dominant drivers of increased flood damages are related to the number of people and extent of development affected by the flood rather than the physical size of the flood.

Clouds

Understanding clouds is scientifically important, because they generate competing feedbacks in the climate system. However, cloud physics are complex, and many important dynamics occur at a spatial resolution finer than those used

in most global climate models, generating substantial uncertainty in the projected changes in cloud cover for many regions of the world.

On the one hand, clouds reflect visible light, so increases in cloud cover, particularly low-altitude cloud cover, can increase the fraction of incoming sunlight that is reflected before it warms the Earth's surface. On the other hand, clouds absorb outgoing infrared radiation leaving the Earth's surface and thus contribute to the greenhouse effect, so increases in cloud cover can amplify warming (Boucher et al. 2013).

To date, global-scale changes in cloudiness remain unclear. Looking forward, some analyses suggest the potential for circulation changes leading to large-scale, nonlinear reductions in low-latitude cloudiness with warming that could substantially increase the sensitivity of temperature to CO_2 forcing (Caballero and Huber 2013).

One way in which cloud science may become important to economists is in the rapidly growing research field of "geoengineering" or "climate engineering," which considers various proposals to intentionally alter the climate so as to counteract some effects of greenhouse gas emissions (Caldeira, Bala, and Cao 2013). The most widely proposed intervention is "solar radiation management," which involves increasing the reflectivity of the atmosphere in order to shade and cool the surface. One proposed mechanism for increasing reflectivity is to spray aerosol precursors into the upper atmosphere, mimicking the mechanism through which historical volcanic eruptions have cooled the surface. Another proposal, with more localized and shorter lasting effects, is "cloud brightening," achieved by manipulating cloud-droplet size by spraying particles into lower portions of the atmosphere. A theoretically appealing feature of cloud-brightening proposals is that cloud brightening might be used to temporarily cool a city or ecosystem during particularly damaging heat waves, although the various economic costs and unintended effects of such policies remain poorly studied (Proctor et al. 2018).

Ocean Acidification

As the CO_2 concentration in the atmosphere increases, some of this CO_2 will dissolve into ocean water, where it will form carbonic acid and increase ocean acidity. Currently, the ocean absorbs roughly one-quarter of global CO_2 emissions through this process.

The rate of acidification depends on local chemistry, temperature, circulation patterns, and freshwater inputs. For example, in the North Pacific during the last three decades, surface ocean acidity has increased by about 12 percent. Globally, the current rate of acidification is unparalleled in at least the last 55 million years, as reflected in a variety of chemical indicators from ocean sediments (Hönisch et al. 2012). In a high-emissions scenario, global mean surface ocean acidity is projected to increase 100–150 percent (Jewett and Romanou 2017).

Ocean acidification is thought to alter marine ecosystems substantially, although the magnitude of these effects is not well understood. The acidity of ocean waters is known to alter the ability of organisms, such as clams or corals, to create the hard shells and reefs that they depend on for survival, effects with largely unknown consequences for the various fish stocks and other marine products consumed around the world.

Ecosystems

Numerous ecosystem changes that can be directly related to climate change have been observed. In many locations around the world, a broad suite of terrestrial organisms is migrating toward higher altitudes and latitudes (Chen, Hill, Ohlemüller, Roy, and Thomas 2011). In the oceans, fish are migrating to stay within their preferred water temperatures (Pinsky, Worm, Fogarty, Sarmiento, and Levin 2013). Under moderate- and high-emissions scenarios, many slow-moving terrestrial species like the coastal redwoods (Roberts and Hamann 2016) may be unable to track the northward movement of climate zones—roughly 0.1–1.3 km per year (Loarie et al. 2009)—quickly enough to stay within their thermal tolerances. In many instances, the ability of species to migrate is further aggravated by fragmentation of habitat. Overall, high-latitude ecosystems are likely to be transformed by invasions from lower latitudes, and extinctions may be common at lower latitudes (Pörtner et al. 2014).

Coral reefs—home to more than a million species—are threatened by both high temperatures and ocean acidification (Hoegh-Guldberg et al. 2007). Bleaching events associated with high temperatures have become more frequent and extensive, with widespread events spanning the tropics in 1998, 2010, and 2015–2016 (Hughes et al. 2017), and a majority of coral reefs around the world are projected to be at risk of degradation even under a low-emissions scenario (Frieler et al. 2013).

The relationships between climate change and ecosystem change can be difficult to untangle, because climate is only one of many human-caused factors affecting ecosystems. For example, land-use change, overexploitation, species introductions, nitrogen deposition, and water resource development also play major roles and may exhibit trends that are correlated with climate change (Chapin et al. 2000). But it is worth noting that although the causes of mass extinction events in Earth's geological history are complex and difficult to pin down, a growing body of evidence suggests that a number of the largest mass extinctions coincided with large-scale climate changes (for example, Payne and Clapham 2012).

Tipping Elements and Critical Thresholds

Nonlinearities and feedbacks in the Earth systems give rise to the potential for multiple stable states of different parts of the Earth system, with potentially rapid lock-in of a state shift once critical thresholds are crossed. These parts of the Earth system are often called "tipping elements," and the their thresholds called "tipping points." However, the "tipping point" language can create confusion about the speed with which state shifts can occur. In popular discourse and much of the economic analysis of climate change, changes associated with a "tipping point" are described as occurring "rapidly." This description is accurate in a geological context, insofar as these changes are rapid relative to other comparable drivers of similar changes in the Earth system. But this description can be misleading in some economic contexts; while some state shifts may be rapid on a human timescale, others may play out over millennia. Below, we summarize a few examples; see Kopp, Shwom, Wagner, and Yuan (2016) for a detailed review.

Tipping elements can exist in the atmosphere/ocean circulation. For example, many of the climate oscillations mentioned earlier, such as El Niño–Southern

Oscillation, occur due to tipping elements, and these patterns may undergo substantial shifts in frequency and amplitude in a warmer climate. Large-scale patterns of ocean circulation, such as the Atlantic Meridional Overturning Circulation—an important component of global ocean circulation that plays a major role in setting temperature, precipitation, and sea level in the North Atlantic—are also potential tipping elements. These atmosphere/ocean tipping elements are among those most likely to undergo rapid shifts.

Tipping elements also occur in ice sheets. For example, positive feedbacks involving ocean-ice sheet interactions might cause sustained ice sheet loss in the Antarctic that would eventually raise global mean sea level by multiple meters or tens of meters. Indeed, some evidence suggests that multiple meters of future sea-level rise from the Antarctic may already be locked in, although depending on the pace of regional warming this rise may take many centuries to manifest.

Tipping elements can also exist in the carbon cycle and in ecosystems. For example, warming of previously frozen soils (permafrost) is allowing microbes to decompose freshly unfrozen organic material into CO_2 and methane. These releases may be an important positive feedback on warming, which could potentially amplify warming by several tenths of a degree in the 21st century. Ecosystems are also well known to undergo rapid regime shifts; coral reefs, whose bleaching is discussed above, are a notable example.

How Economists Can Help Climate Science

We close with a few thoughts on how economists can provide support to climate science.

Improving Emissions Forecasts

Forecasts for greenhouse gas emissions in the coming decades and centuries are a key ingredient to physical simulations of climate change. Emissions clearly depend on global economic activity, but there is not a one-to-one mapping of economic forecasts and the standardized RCP emissions scenarios discussed earlier. Because global emissions are a single time series, there are many possible future configurations of the global economy, technology, and policy that could produce each emissions trajectory.

The coordinated standardization exercise that produced the RCP emissions scenarios also constructed a set of five Shared Socioeconomic Pathways (SSPs), which represent standardized population projections, forecasts of economic growth and convergence, and forecasts of technological change in both the energy sector and adaptation technologies. SSPs can be loosely thought of as potential "states of the world" that might be realized in the future and which no single country can unilaterally change through policy. The narrative for SSP 1 is "Sustainability," representing a world with low barriers to both mitigation and adaptation; SSP 2 represents a "Middle of the Road" scenario; SSP 3 is "Regional Rivalry," representing a world with high barriers to both mitigation and adaptation; SSP 4 is "Inequality," representing a world with high barriers to adaptation but low barriers to mitigation (due to slow economic growth); and SSP 5 is "Fossil-fueled development," representing

a world with high barriers to mitigation but low barriers to adaptation. Different combinations of SSPs and policy choices give rise to different global emission trajectories. For example, SSP 5 in the absence of policy measures can give rise to emissions consistent with RCP 8.5, but other SSPs would require carbon subsidies or similar policies to give rise to such high emissions. In all SSPs, emissions low enough to be consistent with RCPs 4.5 and 2.6 require carbon mitigation policy.

The construction of these Shared Socioeconomic Pathways and the corresponding sets of RCP emissions scenarios represent the output of a modeling program coordinated across numerous research groups (Moss et al. 2010). The energy/agriculture/economic/climate integrated assessment models used to construct the scenarios are elaborate process models that have been assembled by interdisciplinary teams of engineers and economic modelers, mostly from energy economics. At the heart of most models are assumptions about exogenous population growth and about the rate and convergence of technical change. Researchers using these models have addressed many issues over the years: for example, the tradeoffs among different technologies and their roles in meeting different emissions targets (for example, Clarke et al. 2009).

However, many economists with expertise that would be useful to these modeling exercises have remained unengaged with (or unaware of) this enterprise. In our view, deepening engagements with economists in subfields outside of energy economics, such as macroeconomics, development economics, and political economy, will help strengthen and accelerate this research program (see also Barron 2018). Further, there should be a stronger emphasis on using empirical results and hindcasting experiments to constrain the behavior of these models (for example, Calvin, Wise, Kyle, Clarke, and Edmonds 2017).

Focusing Climate Research to Support Investigation of Economic and Social Questions

Much of climate science research is focused on answering key research questions formulated by physical scientists about the nature of the global climate system. These questions are scientifically important and of substantial consequence, but in many cases, key questions or measurements about the climate system that are *economically* or *socially* important remain unanswered. We see three general areas where the potential gains from intellectual exchanges between climate scientists and economists seem large.

First, empirically disentangling the economic consequences of climate change is a large research enterprise (discussed by Auffhammer in this symposium; for a summary, see Carleton and Hsiang 2016). Such analyses universally require some "data engineering" to map physical observations appropriately onto social systems (Auffhammer, Hsiang, Schlenker, and Sobel 2013; Hsiang 2016). Several advances in this literature have arisen from methodological innovations in how physical information from climate science has been summarized and integrated into theoretical and econometric models. For example, Schlenker and Roberts (2009) developed an approach for accounting for the accumulating effects of exposure to extreme heat; Hsiang (2010) developed a method for reconstructing continuous human

exposure to tropical cyclones using standard, albeit limited, meteorological data; Hsiang, Meng, and Cane (2011) introduced a technique for identifying populations heavily impacted by the El Niño–Southern Oscillation; and Proctor et al. (2018) developed an approach to isolate the optical effects of overhead volcanic aerosols. These innovations required both insight into the physics of the climate system coupled to insights from economists regarding the construction of economically meaningful measures. Continued empirical progress will require deepening engagement between researchers in these two fields.

Second, global intercomparison programs for climate change models specify outputs that modeling groups must record in order to participate, but in the past these outputs have been geared towards scientific questions rather than toward calculation of economic outcomes that result from climate change. For example, the expected future correlation of drought events across the major agricultural regions is likely to be important for future food prices, but statistics about such correlated extremes are not commonly computed (Kopp, Easterling, et al. 2017). By supplying the climate modeling community with information on what variables, patterns, and scales are of key economic interest, economists can help the climate modeling community synthesize and output their findings in new and useful ways.

Finally, economists can aid climate scientists in identifying research questions that would be the most valuable to address from a socioeconomic perspective. For example, economists can help distinguish between those economic impacts for which it is most valuable to improve climate model resolution, and those for which is it more important to explore structural or parametric uncertainties, even if at lower resolution. Economists can help distinguish between economic outcomes for which it is more important to constrain long-term climate sensitivity and those for which better characterization of short-term responses, natural variability, or spatial patterns is a priority. Economists can also identify new climate observations that would contribute the greatest value to improved risk-management strategies. In short, economists are well suited to support climate scientists by valuing the different types of information that these scientists could potentially provide to the world.

Climate change management is an urgent and pervasive societal challenge. Natural scientists have had a century's head start over economists in studying the topic, but at this point, input from social scientists can be especially productive. We hope this article can spur more economists to engage in this challenge.

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