Conservation Biology

Special Section

Understanding Recent Climate Change

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Abstract: *The Earth's atmosphere has a natural greenhouse effect, without which the global mean surface temperature would be about 33 °C lower and life would not be possible. Human activities have increased atmospheric concentrations of carbon dioxide, methane, and other gases in trace amounts. This has enhanced the greenhouse effect, resulting in surface warming. Were it not for the partly offsetting effects of increased aerosol concentrations, the increase in global mean surface temperature over the past 100 years would be larger than observed. Continued surface warming through the 21st century is inevitable and will likely have widespread ecological impacts. The magnitude and rate of warming for the global average will be largely dictated by the strength and direction of climate feedbacks, thermal inertia of the oceans, the rate of greenhouse gas emissions, and aerosol concentrations. Because of regional expressions of climate feedbacks, changes in atmospheric circulation, and a suite of other factors, the magnitude and rate of warming and changes in other key climate elements, such as precipitation, will not be uniform across the planet. For example, due to loss of its floating sea-ice cover, the Arctic will warm the most.*

Keywords: climate change, greenhouse effect, radiative forcing, feedbacks, climate sensitivity

Entendiendo el Cambio Climático Reciente

Resumen: *La atmósfera de la Tierra tiene un efecto de invernadero natural, sin el cual la temperatura superficial global promedio sería aproximadamente 33 °C menos y la vida no sería posible. Las actividades humanas han incrementado las concentraciones atmosféricas de bióxido de carbono, metano y otros gases en cantidades traza. Esto ha intensificado el efecto de invernadero, lo que ha resultado en el calentamiento de la superficie. De no ser por los efectos parcialmente compensatorios del incremento de las concentraciones de aerosoles, el incremento en la temperatura superficial global promedio en los últimos 100 años sería mayor que la observada. El calentamiento superficial continuará inevitablemente en el siglo 21 y es muy probable que tenga impactos ecológicos generalizados. La magnitud y tasa de calentamiento del promedio global será determinado principalmente por la fuerza y dirección de las retroalimentaciones climáticas, la inercia térmica de los océanos, la tasa de emisiones de gases de invernadero y las concentraciones de aerosoles. La magnitud y tasa de calentamiento y cambios en otros elementos climáticos clave, como la precipitación, no será uniforme en todo el planeta debido a las expresiones regionales de las retroalimentaciones climáticas, los cambios en la circulación atmosférica y un conjunto de otros factores. Por ejemplo, el Ártico tendrá un mayor incremento de temperatura debido a la pérdida de su cubierta de hielo flotante.*

Palabras Clave: cambio climático, coerción radiante, efecto de invernadero, retroalimentaciones, sensibilidad climática

Introduction

The most fundamental measure of the Earth's climate is the global mean annual surface air temperature (hereafter, global surface temperature). It can be assessed with reasonable reliability back to about 1880 (Fig. 1). The global surface temperature varies considerably from year to year and exhibits decadal-scale variations. This is a reflection of natural climate variability. There is also a general upward trend in global surface temperature. In manifesting warming over the land and ocean parts of both hemispheres, this rise in temperature is global in scope.

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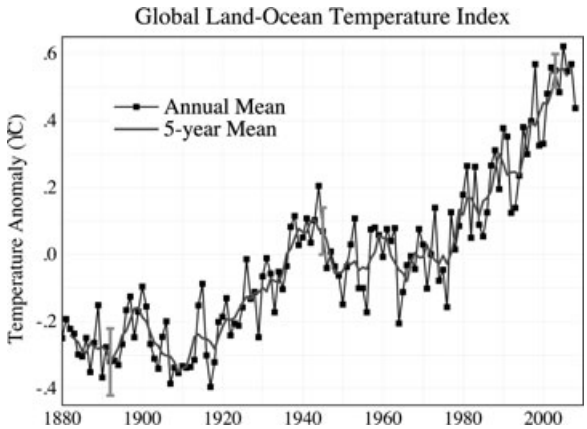

Figure 1. Global surface temperature for 1880 through 2008 on the basis of analyses of the National Aeronautics and Space Administration Goddard Institute for Space Sciences. Error estimates (vertical bars) are shown for selected years (source: http://data.giss.nasa.gov/gistemp/graphs).

The rise in temperature from 1880 through 2008 amounts to about 0.7 °C (Fig. 1). As discussed in numerous peer-reviewed papers and reviewed in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (IPCC-AR4), this has been attended by sea-level rise; reductions in mass of many of the world's glaciers and of the ice sheets of Greenland and Antarctica; a pronounced decline in the extent of the Arctic's floating sea ice; and warming and thawing of permafrost in northern high latitudes. The assembled evidence is now overwhelming that these changes, at least in part, manifest rising concentrations of atmospheric greenhouse gas concentrations, largely due to the burning of fossil fuels. Continued climate change through the 21st century is likely to have major impacts on ecosystems. These may include habitat loss, shifts in vegetation zones, altered ocean chemistry, increased vulnerability to forest fire, infestation of pests and invasive species, and risk to established agricultural productivity.

To help set the stage for this special section of *Conservation Biology*, I provide a brief review of the physical basis of recent climate change and projections for the 21st century. My discussion is built on the IPCC-AR4 (IPCC 2007), classic textbooks, posts in RealClimate (http://www.realclimate.org, an outstanding source of information on climate change), and individual peer-reviewed papers, including some from my own research, which have primarily focused on the Arctic region.

Climate and Climate-Change Theory

Radiative Balance and the Natural Greenhouse Effect

The climate system is powered by the sun. About 99.9% of the electromagnetic radiation emitted by the sun is in wavelengths of 0.15–4 μm (1 μm = 10^{-6} m). In turn, about 50% of the total emitted energy is in the visible part of the spectrum (0.4–0.7 μm). For the globe as a whole, about 30% of the solar radiation received at the top of the atmosphere (TOA) is scattered back to space, primarily by the surface and clouds, but also by atmospheric gases and aerosols (small solid or liquid particles in the atmosphere). About 20% of the incident solar radiation is absorbed within the atmosphere and about 50% is absorbed by the surface. The solar radiation received at the top of the atmosphere minus that which is scattered back to space is known as the net TOA solar flux.

In a steady-state condition, the planet emits as much radiation to space as it absorbs. This is known as radiative balance. Emission to space, however, is in a different denomination, namely long-wave radiation. This emission occurs at wavelengths between 4–300 μm, peaking at about 10 μm (Ahrens 2008).

Imagine the Earth had no atmosphere. In such a fictitious case, the global surface temperature could be computed through application of the Stefan-Boltzmann equation. This fundamental equation states that the emitted radiation from a blackbody (a blackbody absorbs and emits all wavelengths of radiation; the Earth's surface approximates one with respect to long-wave radiation) depends on the fourth power of the temperature (Barry & Chorley 2003). Because radiative balance demands that the planet is radiating just as much energy as it absorbs from the sun, one can plug the absorbed solar flux spread across the globe into the Stefan–Boltzmann equation and solve for the surface temperature. Assuming that, like the real Earth, 30% of the solar flux is scattered back to space, this yields a surface temperature of about −18 °C.

It turns out that the observed global surface temperature is actually about +15 °C, 33 °C higher than computed from the Stefan-Boltzmann equation. The reason for this much higher temperature is that the Earth has an atmosphere that absorbs and emits long-wave radiation. Viewed most simply, some of the long-wave radiation emitted by the surface is absorbed by the overlying atmosphere and reemitted downward. Herein lies the natural greenhouse effect of the planet, without which life would be impossible. The most important of these absorbers and emitters, known as greenhouse gases, is water vapor. It accounts for about 50% of the natural greenhouse effect. Carbon dioxide and other gases such as methane contribute about 25%. Clouds also absorb and emit long-wave radiation and account for the remaining quarter. Unlike a blackbody, however, greenhouse gases absorb and emit in specific wavelengths. For this reason they are called selective absorbers (and emitters). For example, carbon dioxide absorbs and emits strongly at about 15 μm.

The existence of the natural greenhouse effect does not obviate the concept of radiative balance (i.e., the net TOA solar flux balancing the long-wave emission to

space). Nevertheless, given that the observed global surface temperature is much higher than it would be without greenhouse gases, radiative balance in their presence requires that most of the long-wave emission to space comes not from the warm surface, but from the atmosphere, where it is cooler. The emission of long-wave radiation to space is strongest at about 6 km above the surface, roughly in the middle of the troposphere (the layer of the atmosphere closest to the Earth's surface to about 10 km). There is also some direct emission to space from the surface. This occurs at wavelengths at which atmospheric absorption is small.

The concept of radiative balance is only applicable for conditions averaged across the globe. What one actually sees is that from about 38° latitude toward the equator in each hemisphere, more radiation is absorbed by the Earth–atmosphere system than is emitted to space (i.e., there is a radiation surplus). Poleward from this latitude, there is a radiation deficit, with more radiation emitted to space than is absorbed. This results from the circulation of the atmosphere and to a lesser extent the oceans, which transport excess heat from lower latitudes, where the sun's rays strike the Earth strongly, to higher latitudes, where the solar heating is weaker. Trenberth and Caron (2001) provide a valuable review of the phenomenon.

Radiative Forcing

Global radiative balance implies a steady global surface temperature. If the Earth–atmosphere system is thrown out of radiative balance, the surface temperature will change. A key concept in this regard is radiative forcing. Radiative forcing is a measure of the influence that a factor has on disrupting global radiative balance. The larger the forcing is, the larger the disruption of balance. Positive forcing means more radiation coming in than going out. Negative forcing means more radiation going out than coming in. A number of radiative forcings, of different origin and sign, may exist together, and to a first order, can be added to yield a total forcing. If the system is thrown out of radiative balance, it will try to come back into balance, attended by a rise or fall in global surface temperature, depending on whether the radiative forcing is positive or negative, respectively. As part of the process of restoring radiative balance, there will also be temperature changes in the troposphere and stratosphere. Largely as a result of human activities, there now exists a positive radiative forcing consistent with the overall rise in temperature (Fig. 1).

In the framework adopted by the IPCC-AR4, radiative forcings from different sources are estimated for the year 2005, relative to preindustrial conditions, defined as those for the year 1750. They are expressed in terms of watts per square meter (W/m^2), that is, units of energy flux density. Forcings are estimated with the aid of one-dimensional radiative–convective models to assess the instantaneous change in the radiation balance at the top of the tropopause (the layer between the troposphere and stratosphere) to an imposed perturbation, such as that associated with the change in carbon dioxide concentration between preindustrial times (about 280 ppm) and 2005 (about 385 ppm). Although there are good reasons why radiative forcings are assessed at the top of the tropopause instead of the TOA, one can think of them in terms of disruption of the TOA radiation balance without sacrificing too much rigor.

The largest single positive forcing is from carbon dioxide (Fig. 2). With a higher concentration of this gas, there is an imposed addition to the natural greenhouse effect. The atmosphere becomes more opaque to long-wave radiation. More of the long-wave radiation emitted upward from the surface is absorbed by the atmosphere above it, and instead of escaping to space it is directed downward. A key expression of this is that the level at which most of the long-wave radiation escapes to space moves higher into the atmosphere, where it is cooler. Because the atmosphere is radiating from a higher, cooler level, the result is a radiation imbalance, with net TOA solar radiation exceeding the long-wave emission to space. Other greenhouse gases contributing to this process and hence representing positive radiative forcing include methane, halocarbons, and tropospheric ozone.

Black carbon on snow (soot) and solar irradiance are estimated to be small positive forcings. Black carbon on snow reduces the reflectivity (the albedo) of the surface. With a lower albedo, more solar energy is absorbed. The irradiance forcing means a small increase in solar output, which unlike the other radiative forcings listed is of natural origin. Changes in land use have acted to increase the surface albedo, representing a small negative forcing.

The largest negative forcings are associated with aerosols (primarily sulfate, organic carbon, black carbon, nitrate, and dust). Aerosol effects can be divided into direct effects, associated with scattering and absorbtion of solar radiation in the atmosphere (meaning less energy at the surface, hence cooling), and indirect effects, largely associated with the effects of aerosol loading in increasing the albedo of clouds. Although aerosols have countered some of the positive radiative forcing associated with increased carbon dioxide and other greenhouse gases, the magnitude of aerosol forcings is still quite uncertain. Unlike the major greenhouse gases that are well mixed through the atmosphere (they have approximately the same concentrations across the globe), aerosol effects can also operate on smaller scales. In other words, although one can estimate global radiative forcing associated with aerosols, this global value masks large spatial variations. Aerosol loading comes primarily from fossil–fuel burning (Fig. 2). Large volcanic eruptions inject sulfur gases into the stratosphere that then form aerosols, which can be unambiguously associated with periods of climate cooling lasting a couple of years.

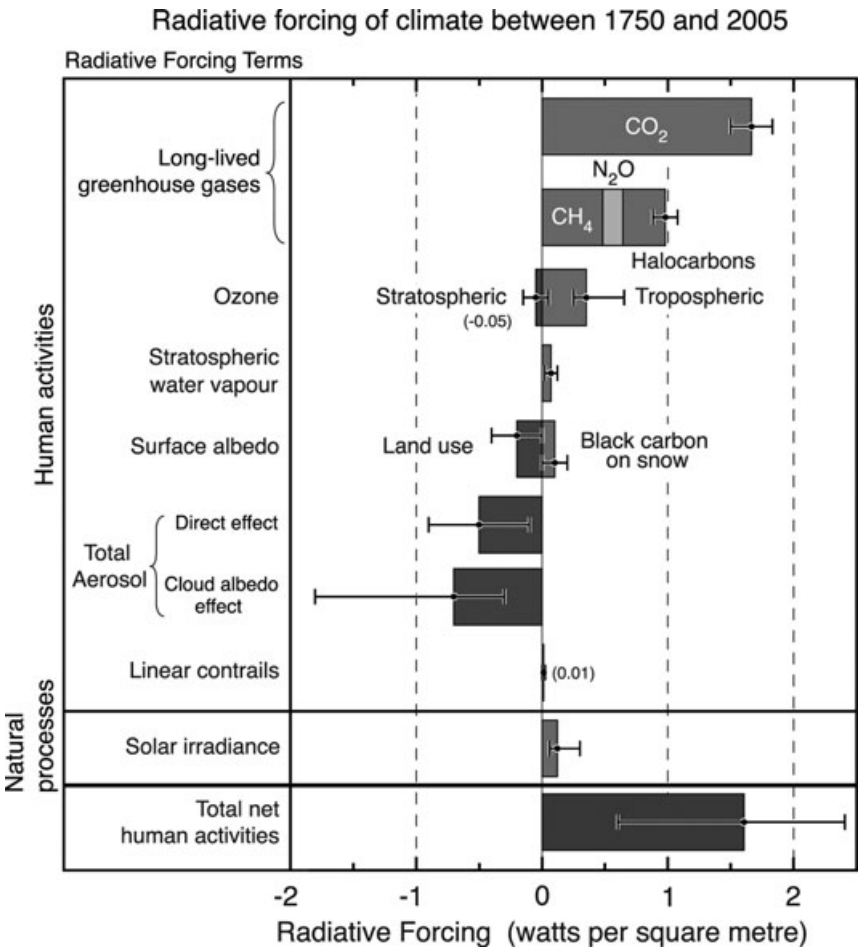

Figure 2. Estimates of global radiative forcing in 2005. The range of estimates is shown for each forcing component and for the total anthropogenic forcing. Volcanic forcing is not shown because of its episodic nature (Source: Climate Change 2007).

Volcanic eruptions, however, are episodic (the last major event was Mt. Pinatubo in 1991), so their forcing is not included.

Summing all the anthropogenic components together, there is a global radiative forcing of about 1.6 W/m^2. Most of the uncertainty in this value is due to the uncertainty in aerosol effects.

Climate Sensitivity and Feedbacks

How much does the global mean surface temperature change in response to a radiative forcing of a given magnitude (or net magnitude of a set of forcings in combination)? How long does it take for the change to occur? These are among the most pressing questions in climate-change science.

The first question deals with the issue of climate sensitivity. Classically, climate sensitivity is viewed as the equilibrium change in global surface temperature in response to a radiative forcing. For example, if more carbon dioxide is added to the atmosphere, which throws the system out of radiative balance, the surface and atmosphere would warm up and start to emit more long-wave radiation to space. Eventually, radiative balance would be restored again. The best current estimate of equilibrium climate sensitivity is about a 0.75 °C change in global surface temperature for every watt per square meter of forcing. This estimation is based on results from climate models and evidence from paleoclimate records. The idea of equilibrium climate sensitivity is complicated by many factors. There are different ways of defining radiative forcings. In addition, not all forcings are equal. Although to a first order one can simply add different forcings together, in reality they have different "efficiencies" in terms of their warming potential (Hansen et al. 2005*b*) Furthermore, because forcings are always changing, equilibrium is a constantly moving target and is never really achieved. Nevertheless, equilibrium climate sensitivity has proven an extremely valuable framework.

The key determiner of the equilibrium climate sensitivity is the sign and strength of the fast feedbacks in the climate system (i.e., those that operate on short timescales, days to decades). Positive feedbacks increase the equilibrium temperature change associated with a forcing, whereas negative feedbacks dampen the change.

The most important of these fast feedbacks is associated with water vapor. Warming initiated by a radiative forcing fosters more evaporation, and a warmer atmosphere can carry more water vapor. Nevertheless, because water vapor is also a greenhouse gas, it causes further warming (Soden et al. 2002). Another is the albedo

feedback associated with snow cover and sea ice (Serreze & Francis 2006). If the temperature rises, some of the Earth's snow cover and sea ice melts, reducing the surface albedo of the planet. With a lower albedo, more solar energy is absorbed at the surface, causing further warming. These are both positive feedbacks because they amplify the global mean change in surface temperature initiated by the forcing. If one were to somehow reduce the carbon dioxide concentration in the atmosphere, thereby imposing a negative climate forcing, the positive feedbacks would foster further cooling. The climate system also has slow feedbacks, such as those associated with changes in the carbon cycle and volume of ice sheets operating over many thousands of years. Although these slow feedbacks are key for understanding the great ice ages of the Pleistocene (Covey et al. 1996), the key to recent climate change is the fast feedbacks.

A fascinating feature of the global climate system is that positive feedbacks dominate. For example, one of the responses to increasing greenhouse gas concentrations could be an increase in cloud cover, which by increasing the albedo of the planet would represent a negative (and fast) feedback. Nevertheless, this and other potential negative feedbacks only appear to be capable of slowing the rate of warming.

It appears that the bulk of the response to a radiative forcing with the fast feedbacks at work occurs over a time span of 20–50 years (Hansen et al. 1985). Much of this deals with the large thermal inertia of the oceans. The basic issue is that the oceans can absorb and store a great deal of heat without a large rise in the surface temperature, keeping the system out of radiative balance. Eventually, with enough heat gain, the radiating surface temperature will become high enough for radiative balance to be restored, but it takes a while for the adjustment to occur. Consider what is expected, given the current radiative forcing of 1.6 W/m^2. If the equilibrium climate sensitivity is 0.75, then this radiative forcing will yield about 1.2 °C of warming. Over the instrumental record, the global mean temperature has risen by about 0.7 °C. This implies there is about another 0.5 °C that has yet to be realized, often referred to as the warming "in the pipeline." This should roughly equate to the current TOA radiation imbalance. As it turns out, the heat in the pipeline, estimated on the basis of the assumed climate sensitivity and the observed global temperature rise, agrees rather well with the value estimated independent from observed changes in ocean heat content over the last decade or so (Hansen et al. 2005*a*).

Projections for the 21st Century

Approaches and Uncertainties

Projections of climate change through the 21st century rely strongly on simulations with coupled global climate models and assumptions about future radiative forcings. Global climate models couple an atmospheric model with models of ocean circulation, land surface, and sea ice (McGuffie & Henderson-Sellers 2005). Modeling teams across the world have provided simulations for use in the IPCC-AR4. During model development and refinement, much attention is given to the ability of a model to simulate the present climate state. This includes comparison of patterns of atmospheric and ocean circulation, temperature, and precipitation as simulated by the model with available observations. Projections of climate change through the 21st century typically involve evaluating how the modeled system evolves with an assumed evolution of changes in concentrations of greenhouse gases and other forcings. Climate elements, such as temperature and precipitation, at different times through the simulation are typically expressed as anomalies with respect to a chosen baseline period taken to represent late 20th century climate (e.g., the period 1980–1999).

Models vary in terms of horizontal and vertical resolution and level of complexity in treating different physical processes. This contributes to scatter among models in their depictions of present-day and projected future climate states. The approach typically taken is to consider the multimodel average (often termed the ensemble average) and the spread between simulations from each model. One finding with this approach is that if one conducts simulations using the best estimates of forcings through the 20th century (both natural and anthropogenic), the ensemble average time series of global temperatures matches up rather well to observations. Although valuable, this approach has shortcomings related to uncertainty in both radiative forcings and modeled climate sensitivity (Kiehl 2007). A related strategy is to examine the mean and spread of multiple simulations from the same model, with each simulation (ensemble member) starting from slightly different initial conditions. This approach recognizes that natural climate variability is as much a part of the model world as it is in the real world. For example, imagine that one had six simulations from which one compiled six maps of temperature anomalies across the globe for some given 20-year average in the 21st century. Because the model could be in a different phase of its own natural variability in each simulation, the six map patterns could look quite different. Averaging results from the six simulations averages out this noise, allowing the greenhouse-gas signal to emerge.

Although climate models are constantly being improved to make better projections of climate, a fundamental unknown is human behavior. Asking how much global temperature will rise during the 21st century is not particularly meaningful. The question is better phrased in terms of how much the temperature will change with given assumed growth rates of greenhouse gases and aerosols. In recognition of this, the IPCC has spent a great deal of effort on coming up with various emissions

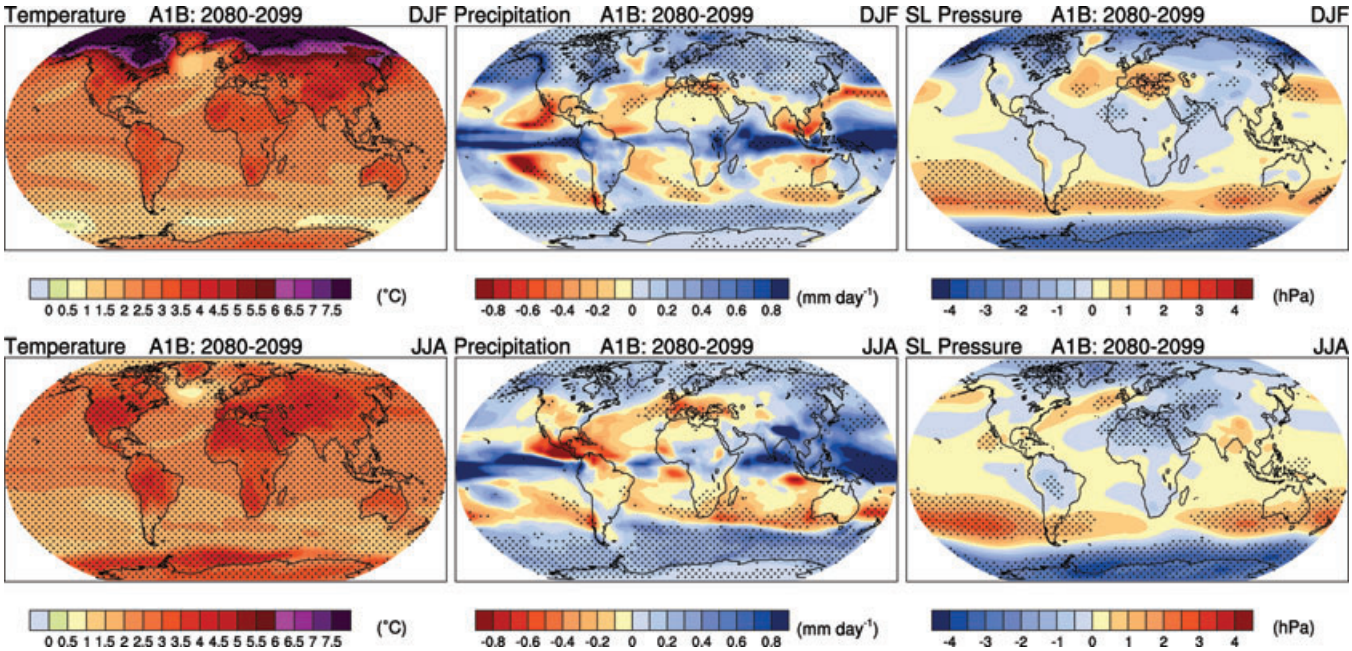

Figure 3. Projected changes for winter (DJF) and summer (JJA) of surface temperature, precipitation, and sea-level (SL) pressure for 2080–2099, relative to 1980–1999. Results shown are average output from simulations with a series of models using the A1B emissions scenario (Source: Climate Change 2007).

scenarios. These are detailed in the *Special Report on Emission Scenarios* (Nakicenovic et al. 2000). They are based on different models of economic growth, population growth, and assumptions regarding the adoption of new technologies. Much of the focus has been on the B1, A1B, and A2 scenarios, which range from modest greenhouse gas growth (B1) to aggressive growth (A2). Of these, the middle-of-the-road A1B scenario has been the most widely adopted. In this scenario, carbon dioxide reaches a level of 720 ppm by 2100 (the current value is about 387 ppm).

Projections

For the multimodel average with the A1B emissions scenario, the global surface temperature in the year 2100, with respect to the base period 1980–1999, is projected to increase by about 2.8 °C. Considering the spread between different models, however, the rise could be as small as 2.2 °C or (because the probability distribution from the simulations is right tailed) larger than 3.0 °C. This spread in projected temperature change implies that different models have different climate sensitivities. With the more modest B1 scenario, the multimodel average change is 1.8 °C. Assuming the aggressive A2 scenario, the multimodel average change is 3.6 °C. These projections do not account for carbon-cycle feedbacks, which could contribute, perhaps substantially, to atmospheric greenhouse gas concentrations and hence additional warming. One of these feedbacks is acidification of the oceans, which leads to reduced capacity of the oceans to absorb part of the anthropogenic carbon emissions. Ocean acidification is also of growing concern with respect to impacts on coral reefs and other marine calcifiers (Dodge et al. 2008). Another potential feedback involves release of carbon locked up in Arctic and subarctic soils. This occurs through increased soil microbial activity as permafrost thaws (Dutta et al. 2006). It is a good example of a biological feedback on climate.

Multimodel mean changes in spatial patterns of temperature, precipitation, and sea-level pressure demonstrate several features (Fig. 3). The maps are depicted as changes averaged over the period 2080–2099 relative to the period 1980–1999 for the A1B scenario. Although recognizing that there is large scatter between individual models in spatial patterns of change, some clear patterns emerge.

The most obvious feature of the winter temperature map is that the largest warming is projected to occur in the Arctic. This feature, known as Arctic amplification (Holland & Bitz 2003; Serreze & Francis 2006), appears in simulations from even early-generation global climate models (e.g., Manabe & Stouffer 1980). The enhanced warming is focused over the Arctic Ocean during the cold season. As the climate warms, the summer melt season lengthens and intensifies, leading to less sea ice at summer's end. Summertime absorption of solar energy in expanding, dark, open-water areas will increase the heat content of the ocean. Ice formation in autumn and winter, important for insulating the warm ocean from the cooling atmosphere, is delayed. This promotes enhanced upward heat fluxes, seen as strong warming at the surface

and in the lower troposphere. This is part of the fast albedo feedback. Areas of open water also represent moisture sources, which can augment warming through the water–vapor feedback. As is also evident, Arctic amplification is not prominent in summer itself, when energy is used to melt remaining sea ice and increase the heat content of the upper ocean, limiting changes in surface temperatures. Loss of snow cover contributes to an amplified temperature response over northern land areas (another component of the albedo feedback), but it is not as pronounced as over the ocean (Serreze & Francis 2006).

Projected changes in precipitation also show a fairly coherent, albeit complex, pattern. This is a reflection of the increased moisture-holding capacity of the warmer atmosphere and changes in atmospheric circulation. In areas where there are efficient precipitation-generating mechanisms, precipitation is expected to increase (areas near the equator, where there is strong convection, and in middle and higher latitudes, where ascent and cooling of poleward-moving airmasses allows water vapor to condense). Sea-level pressure is expected to fall in the high latitudes, with compensating increases in parts of the middle latitudes and subtropics, where precipitation amounts are expected to decline. This follows because high pressures at the surface tend to be accompanied by descending air motion unfavorable to condensation.

Discussion

One can think of recent global climate change in terms of a three-step process of radiative forcing, feedback, and response. Response in this context is the change in the global surface temperature. The key to how much the surface temperature changes for a forcing of given magnitude (i.e., the climate sensitivity) depends on the climate feedbacks. For climate change over the next century, the dominant ones are the positive water vapor and albedo feedbacks, but carbon-cycle feedbacks may well become very important. Radiative forcings will continue to evolve. This evolution, in concert with delayed responses (largely due to the thermal inertia of the oceans), means the system will tend to remain out of equilibrium. Because of regional expressions of climate feedbacks, changes in atmospheric circulation and a suite of other factors, the magnitude and rate of warming and changes in other key climate elements, such as precipitation, will not be uniform across the planet. Although projections of regional change are the most uncertain aspects of climate-model simulations, there is near-universal agreement that the Arctic will warm the most.

Distinction between forcing and feedback as viewed above applies specifically to the framework of changes in global surface temperature. It may not apply in a framework of regional change. For example, regarding its role in boosting change in global surface temperature in response to radiative forcing from increased greenhouse gas concentrations, the process of Arctic amplification associated with the loss of sea ice is clearly a feedback. In a study of processes contributing to temperature change specifically over the Arctic Ocean, however, one may correctly consider loss of sea ice as a forcing on temperature. Hence, when discussing forcings and feedbacks, one must be aware of the framework being considered.

An emerging issue is that in some respects, observed changes are outstripping expectations. The Arctic's shrinking sea-ice cover is perhaps the best example. From satellite remote sensing, accurate measurements of the extent of sea ice are available for 1979 to the present. Regression analysis documents downward trends in ice extent for every month, largest in September, the end of the melt season (Serreze et al. 2007). Extreme September extent minima have characterized every year since 2002, with September 2007 setting a new record (Stroeve et al. 2008) that was nearly matched by September 2008 (http://nsidc.com/arcticseaicenews/). Extent for September 2009 was the third lowest on record. Zhang and Walsh (2006) note that essentially all models considered in the IPCC-AR4 show declining sea ice over the period of observations. Although this is strong evidence of a role of increased greenhouse gases on the observed trend, Stroeve et al. (2007) find that, depending on the time window for analysis, none or few simulations have a September trend as large as observed. One explanation is that natural variability in the observed Arctic system has been a very strong player. Changes in spring cloud cover, wind-driven alterations in sea-ice circulation and ice age, and altered ocean heat transport have all been implicated in the observed ice retreat. An alternative explanation, however, is that the models, as a group, underrepresent sea-ice sensitivity to greenhouse gas loading (Stroeve et al. 2007).

Could the summer sea-ice cover be lost as early as 2030? This no longer seems unreasonable, and there are some in the community who think this may be an optimistic assessment. This does not bode well for species such as the polar bear, and there is evidence that impacts of the loss of sea ice are already cascading through the marine food chain of the northern Bering Sea (Grebmeier et al. 2006). Where some of this thinking comes from is growing recognition that as the ice thins, it can become vulnerable to a "kick" associated with natural variability, leading to rapid loss of the remaining ice cover through the albedo feedback process (Holland et al. 2006). The summer of 2007 exhibited a pattern of atmospheric circulation seemingly perfect for melting sea ice. Could this have been the kick that initiates further rapid ice loss? This is the sort of question that can be answered only in hindsight.

Acknowledgments

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