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METEO 469 Course Outline

Lesson 1- Introduction to Climate and Climate Change

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

Credit: Today in Montclair, 94611 [1]

Did you complete the Course Orientation?

Before you begin this course, please complete the Course Orientation.

About Lesson 1

Human-caused climate change represents one of the great environmental challenges of our time. To appreciate its societal, environmental, and economic implications, one must appreciate the basic underlying science. This course seeks to first lay down the fundamental scientific principles behind climate change and global warming. These principles involve aspects of atmospheric science and meteorology, as well as aspects of other areas of the physical and biological sciences. With a firm grounding in the basic science, we go on to explore other issues involving climate change impacts and the issue of *mitigation* — that is, solutions to dealing with the challenges presented by climate change.

In the process, we will learn how to do basic computations and to use theoretical models of the climate system of varying complexity to address questions regarding future climate change. Students will explore the impacts of various alternative greenhouse gas emissions scenarios and investigate policies that would allow for appropriate stabilization of future greenhouse gas concentrations. The structure of the course roughly parallels the treatment of the subject matter by the reports of the *Intergovernmental Panel on Climate Change* (*IPCC*), focusing first on the basic science, then the future projections and their potential impacts, and finally issues involving adaptation, vulnerability, and mitigation. We will use a variety of tools to inform our understanding of these topics, including digital video, audio, simulation models, and virtual field trips to online data resources.

In this first lesson, we are going to define *climate* and *climate change*, as well as the closely related matter of *global warming*. We will introduce the components of the climate system (the atmosphere, ocean, cryosphere, and biosphere). We will also briefly introduce some of the other key scientific concepts: atmospheric structure and composition, energy balance, atmospheric and oceanic circulation. We will draw a distinction between natural and human impacts on climate, and we will review the science behind greenhouse gases and the greenhouse effect. We will explore the crucial topic of feedback mechanisms, including important emerging knowledge regarding *carbon cycle feedbacks*. Finally, we will begin to explore a number of important overriding themes such as the role of scientific uncertainty in decision making.

What will we learn in Lesson 1?

By the end of Lesson 1, you should be able to:

- define the Earth's climate system and its components;
- distinguish the factors governing natural climate variability from human-caused climate change;
- explain the greenhouse effect;
- describe the role of feedback mechanisms; and
- discuss the role of uncertainty in decision-making.

What will be due for Lesson 1?

Please refer to the Syllabus for the specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 1. Detailed directions and submission instructions are located within this lesson.

- Read:
  - IPCC Fifth Assessment Report, Working Group 1 [2]
  - Summary for Policy Makers [3]
    - A Introduction: p. 4
    - B.5 Observed Changes in the Climate System, Carbon and Other Biogeochemical Cycles: p. 11-12
    - C Drivers of Climate Change: p. 13-14
  - Dire Predictions, v.2: p. 8, 13, 22-25
- Participate in the *LESSON 1: GENERAL DISCUSSION OF METEO 469* discussion forum.

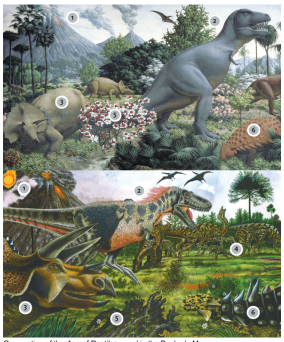
What is Climate?

When it comes to defining climate, it is often said that "climate is what you expect; weather is what you get". That is to say, climate is the statistically-averaged behavior of the weather. In reality, it is a bit more complicated than that, as climate involves not just the atmosphere, but the behavior of the entire climate system—the complex system defined by the coupling of the atmosphere, oceans, ice sheets, and biosphere.

Having defined climate, we can begin to define what climate change means. While the notion of climate is based on some sort of statistical average of behavior of the atmosphere, oceans, etc., this average behavior can change over time. That is to say, what you "expect" of the weather is not always the same. For example, during El Niño years, we expect it to be wetter in the winter in California and snowier in the southeastern U.S., and we expect fewer tropical storms to form in the Atlantic during the hurricane season. So, climate itself varies over time.

If climate is always changing, then is *climate change* by definition always occurring? Yes and No. A hundred million years ago, during the early part of the Cretaceous period, dinosaurs roamed a world that was almost certainly warmer than today. The geological evidence suggests, for example, that there was no ice even at the North and South poles. So global warming can happen naturally, right? Certainly, but *why* was the Earth warmer at that time?

A hint of *why* can be found in many of the careful renditions of what the Earth may have looked like during the age of dinosaurs. Some of the most insightful interpretations came from the 19th century Yale paleontologist, Othniel Charles Marsh. Let us look at one of his renderings:


One section of the *Age of Reptiles* mural in the Peabody Museum
Credit: Yale Alumni Magazine [4]

Think About It!

What features of the above mural might provide a clue for why the early Cretaceous was so warm?

Click for answer.

So, the major climate changes in Earth's geologic past were closely tied to changes in the greenhouse effect. Those changes were natural. The changes in greenhouse gas concentrations that we talk about today, are, however, not natural. They are due to human activity.

Importance: Why Should We Care About Climate Change?


Figure 1.1: Polar bears walking on melting ice.
Credit: The Select Committee on Energy Independence and Global Warming [5]

As we have discussed, climate change can be natural. If climate changes naturally, then why should we be concerned about the climate change taking place today? After all, the early Cretaceous period discussed previously was warmer than today, but life thrived even in regions, such as the interior of Antarctica, that are uninhabitable today.

One misconception is that the threat of climate change has to do with the absolute warmth of the Earth. That is not, in fact, the case. It is, instead, the *rate* of change that has scientists concerned. Living things, including humans, can easily adapt to substantial changes in climate as long as the changes take place slowly, over many thousands of years or longer. However, adapting to changes that are taking place on timescales of *decades* is far more challenging.

Here is a useful "thought experiment" to illustrate what sort of discussion might be happening now if, instead of the current climate, we were living under the climate conditions of the last Ice Age, and human fossil fuel emissions were pushing us out of the ice age and into conditions resembling the pre-industrial period, rather than the actual case, where we are pushing the Earth out of the pre-industrial period and into a period with conditions more like the Cretaceous. Take a look at Figure 1.2 below, which indicates the Gulf coast continental outline near the height of the last Ice Age 18,000 years ago, vs. the current continental outline.

[6]
Figure 1.2: Sea Level: Today vs. the Pleistocene Epoch - *Enlarge* [6]
Credit: EPA [7]

Everything in the the lighter shading would be flooded in the transition from the ice age to pre-industrial modern climate. But what sort of effort would that have taken?

It turns out that the natural increase in atmospheric CO_2 that led to the thaw after the last Ice Age was an increase from 180 parts per million (ppm) to about 280 ppm. This was a smaller increase than the present-time increase due to human activities, such as fossil fuel burning, which thus far have raised CO_2 levels from the pre-industrial value of 280 ppm to a current level of about 400 pm--a level which is increasing by 2 ppm every year. So, arguably, if the dawn of industrialization had occurred 18,000 years ago, we may very likely have sent the climate from an ice age into the modern pre-industrial state.

How long it would have taken to melt all of the ice is not precisely known, but it is conceivable it could have happened over a period as short as two centuries. The area ultimately flooded would be considerably larger than that currently projected to flood due to the human-caused elevation of CO_2 that has taken place so far. The hypothetical city of "Old Orleans" would have to be relocated from its position in the Gulf of Mexico 100+ miles off the coast of New Orleans, to the current location of "New Orleans".

By some measures, human interference with the climate back then, had it been possible, would have been even more disruptive than the current interference with our climate. Yet that interference would simply be raising global mean temperatures from those of the last Ice Age to those that prevailed in modern times prior to industrialization. What this thought experiment tells us is that the issue is not whether some particular climate is objectively "optimal". The issue is that human civilization, natural ecosystems, and our environment are heavily adapted to a particular climate — in our case, the current climate. Rapid departures from that climate would likely exceed the adaptive capacity that we and other living things possess, and cause significant consequent disruption in our world.

Here is an example that relates how polar bears in the north will have to adapt to the changing climate conditions that are presenting themselves in northern Manitoba, from a panel discussion shot in November, 2010 including the designer of this course and the popular star George Stroumboulopoulos of CBC Television in Canada (Michael Mann's comments begin around 4:48 in the linked video [8]):


Figure 1.3: CBC's Stroumboulopoulos Polar Bear Special (course author on the right).

So, hopefully, we have established that climate change is something worth caring about. Perhaps it is something worth *doing* something about. But you cannot really do anything about a problem that you do not understand, let alone know how to solve.

So, in the remainder of this lesson, we are going to try to begin to get a handle on the fundamental science underlying climate change and global warming.

Overview of the Climate System - Part 1

The components of the climate system

The climate system reflects an interaction between a number of critical sub-systems or *components*. In this course, we will focus on the components most relative to modern climate change: the atmosphere, hydrosphere, cryosphere, and biosphere. Please click on the arrow in the screen-cast below to walk through the important aspects of these components.

Credit: Schematic of Climate System, IPCC 2007 [9]

Atmospheric Structure and Composition

The atmosphere is of course a critical component of the climate system, and the one we will spend the most time talking about.

One key feature about the atmosphere is the fact that pressure and density decay exponentially with altitude:

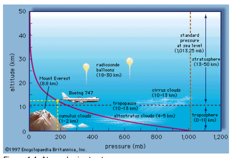


Figure 1.4: Atmospheric structure.
Credit: Britannica.com

As you can see, the pressure decays nearly to zero by the time we get to 50 km. For this reason, the Earth's atmosphere, as noted further in the discussion below, constitutes a very thin *shell* around the Earth.

The exponential decay of pressure with altitude follows from a combination of two very basic physical principles. The first physical principle is **the ideal gas law**. You are probably most familiar with the form $pV = nRT$, but that form applies to a bounded gas, where the volume can be defined. In our case, the gas is free, and the appropriate form of the ideal gas law is

$$p = \rho RT$$

(1)

where p is the atmospheric pressure, ρ is the density of the atmosphere, $R = 287 \; J \, K^{-1} \, kg^{-1}$ is the gas constant that is specific to Earth's atmosphere, and T is temperature.

The 2nd principle is the **force balance**. There are two primary vertical forces acting on the atmosphere. The first is gravity, while the other is what is known as the pressure gradient force — it is the support of one part of the atmosphere acting on some other part of the atmosphere. This balance is known as the *hydrostatic balance*.

The relevant pressure gradient force in this case is the *vertical* pressure gradient force. When we are talking about a continuous fluid (which the atmosphere or ocean is), then the correct form of force balance involves force per unit volume of fluid.

In this form, we have for gravity (the negative sign indicates a downward force):

$$F_{gravity} = -\rho \times g$$

(2)

where g is Earth's surface gravitational acceleration ($9.81 \text{ meters/second}^2$).
The pressure gradient force has to be written in terms of a *derivative*:

$$F_{pgf} = {dp}/{dz}$$

(3)

The positive sign insures that an atmosphere with greater density below exerts a positive (upward) force.

In equilibrium, these forces must balance, i.e.

$${dp}/{dz} = -\rho \times g$$

(4)

Now we can use the ideal gas law (eq. 1.) to substitute for ρ, the expression $\rho = p/RT$, giving

$${dp}/{dz} = {-p}/{RT} \times g$$

(5)

or re-arranging a bit,

$${dp}/{p} = -\left({g}/{RT}\right) dz$$

(6)

The term in parentheses can be treated as a constant (in reality, temperature varies with altitude, but it varies less dramatically than pressure or density, so it's easiest to simply treat it as a constant).

This is a relatively simple first order differential equation.

Self Check...

Do you remember how to solve this first order differential equation from your previous math studies?

Click for answer.

The expression for atmospheric pressure as a function of altitude is:

$$p = p_0 \exp\left[-\left({g}/{RT}\right)(z - z_0)\right]$$

(7)

where p_0 is the surface pressure, and z_0 is the surface height (by convention typically taken as zero).

This equation is known as the *hypsometric equation*.

The combination ${g}/{RT}$ has units of inverse length, and so we can define a *scale height* (assuming a mean temperature $T = 14^{\circ}C = 287K$,

$$h_s = {RT}/{g} \approx 8.4km$$

(8)

and write:

$$p/p_0 = \exp\left[-\left(z - z_0\right)/h_s\right]$$

(9)

this gives an exponential decline of pressure with height, with the *e-folding height* equal to the *scale height*, representing the altitude at which pressure falls to roughly 1/3 of its surface value. At this altitude, which as you can see from the above graphic is just a bit below the height of Mt. Everest, roughly 2/3 of the atmosphere is below you.

Self-check...

Using the hypsometric equation (9 above), estimate the altitude at which roughly half of the atmosphere is below you.

Click for answer.
Let us look at the vertical structure of the atmosphere in more detail, define some key layers of the atmosphere:

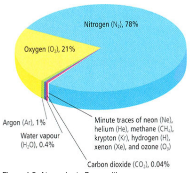


Figure 1.5: Atmospheric Composition.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*

Now, let us talk a bit more about the atmospheric *composition*:

The atmosphere is mostly nitrogen and oxygen, with trace amounts of other gases. Most atmospheric constituents are *well mixed*, which is to say, these constituents vary in constant relative proportion, owing to the influence of mixing and turbulence in the atmosphere. The assumption of a *well-mixed atmosphere* and the assumption of *ideal gas behavior*, were both implicit in our earlier derivation of the exponential relationship of pressure with height in the atmosphere.

There are, of course, exceptions to these assumptions. As discussed earlier, ozone is primarily found in the lower stratosphere (though some is produced near the surface as consequence of photochemical smog). Some gases, such as methane, have strong *sources* and *sinks* and are therefore highly variable as a function of region and season.

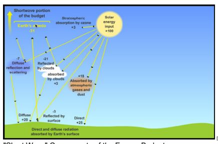
Atmospheric water vapor is highly variable in its concentration, and, in fact, undergoes phase transitions between solid, liquid, and solid form during normal atmospheric processes (i.e., evaporation from the surface, and condensation in the form of precipitation as rainfall or snow). The existence of such phase transitions in the water vapor component of the atmosphere is an obvious violation of ideal gas behavior!

Of particular significance in considerations of atmospheric composition are the so-called greenhouse gases (CO_2, water vapor, methane, and a number of other trace gases) because of their radiative properties and, specifically, their role in the so-called *greenhouse effect*. This topic is explored in greater detail later on in this lesson.

Overview of the Climate System (part 2)

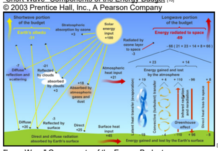
Basics of Energy Balance and the Greenhouse Effect

An interactive animation provided below allows you to explore the balance of incoming and outgoing sources of energy within the climate system. A brief tutorial is provided below, first with the *short wave* component and then the *long wave* component of the energy budget. (Click image or link below to open the animation in a new window.)



[10]

"Short Wave" Components of the Energy Budget [10]



[11]

"Long Wave" Components of the Energy Budget
[Please note that there is a slight error in the spoken part of the "Long Wave" video above, beginning around 3:17. While adding up the components of the long wave energy emitted by the atmosphere to space, which total 66 units, the narrator neglects to add in 8 units of direct heat loss to space.]

Now explore this tool yourself, at your own pace. It takes some time to absorb all of the information that is contained here. Start with the short wave energy budget. Once you are satisfied that you have got that down, go on to the somewhat more complex long wave energy budget.

Short wave Energy Budget

Consider how incoming and outcoming energy sources of shortwave and longwave radiation achieve a net balance:

- at the surface
- within the atmosphere
- at the top of the atmosphere

In future lessons, we will examine the *greenhouse effect* in a more *quantitative* manner. Note here how the *greenhouse effect* works qualitatively. It involves the ability of *greenhouse gases* within the atmosphere to absorb longwave radiation, impeding the escape of the longwave radiation emitted from the surface to outer space.

In our first discussion session at the end of this lesson, you will be asked to speculate on certain aspects of this schematic, and to pose some questions of your own for your classmates to attempt to answer.

Seasonal and Latitudinal Dependence of Energy Balance

Next, let us note that the above picture represents *average climate conditions*, that is, averaged over the entire Earth's surface, and averaged over time. However, in reality, the incoming distribution of radiation varies in both space and time. We measure the radiation in terms of power (energy per unit time) per unit area, a quantity we term *intensity* or *energy flux*, which can be measured in watts per square meter (W/m^2).

The dominant spatial variation occurs with latitude. On average, there is roughly 343 W/m^2 of incoming short wave solar radiation that is incident on the Earth, averaged over time, and over the Earth surface area. Obviously, there is more incoming solar radiation arriving at the surface near the equator than near the poles. On average, roughly 30%, or about 100 W/m^2 of this incident radiation is reflected out to space by clouds and reflective surfaces of the Earth, such as ice and desert sand, leaving roughly 70% of the incoming solar radiation to be absorbed by the Earth's surface. The portion that is reflected by clouds and by the surface also varies substantially with latitude, owing to the latitudinal variations in cloud and ice cover:

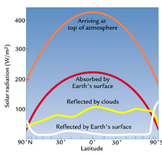


Figure 1.6: Latitudinal Distribution of Various Sources of Incoming and Outgoing Radiation
Credit: Ruddiman, *Earth's Climate: Past and Future* (W.H. Freeman, 2001)

Moreover, the distribution of outgoing long wave radiation also varies substantially with latitude:

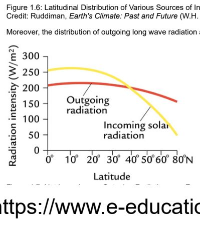


Figure 1.7: Net Incoming vs. Outgoing Radiation as a Function of Latitude
Credit: Ruddiman, *Earth's Climate: Past and Future* (W.H. Freeman, 2001)

More terrestrial radiation is emitted from the warmer tropical regions and less emitted from the cold polar regions:

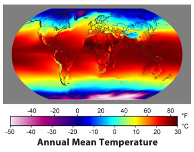


Figure 1.8: Annual Mean Temperature
Credit: Wikimedia Commons

The disparity shown above (Figure 1.8) between the incoming solar radiation that is absorbed at the surface and the outgoing terrestrial radiation emitted from the surface poses a conundrum. As we can see in Figure 1.8, outgoing radiation exceeds incoming radiation near the poles, i.e., there is a deficit of radiation at the surface. Conversely, there is a surplus of incoming radiation near the equator. Should the poles, therefore, continue to cool down and the tropics continue to warm up over time?

Think About It!

Any idea what the solution to this conundrum might be?

Click for answer.

It is also worth noting that the incoming solar radiation is not constant in time. As we will see in later lessons, the output of the Sun, the so-called solar constant, can vary by small amounts on timescales of decades and longer. During the Earth's early evolution, billions of year ago, the Sun was probably about 30% less bright than it is today--indeed, explaining how the Earth's climate could have been warm enough to support life back then remains remains somewhat of a challenge, known as the *"Faint Young Sun" paradox* [12].

Even more dramatic changes in solar insolation take place on shorter timescales—the diurnal and annual timescale. These changes, however, do not have to do with the net output of the Sun, but rather the distribution of solar insolation over the Earth's surface. This distribution is influenced by the Earth's daily rotation about its axis, which of course leads to night and day, and the annual orbit of the Earth about the Sun, which leads to our seasons. While there is a small component of the seasonality associated with changes in the the Earth-Sun distance during the course of the Earth's annual orbit about the Sun (because of the slightly elliptical nature of the orbit), the primary reason for the seasons is the tilt of Earth's rotation axis relative to the plane defined by the Earth and the Sun, which causes the Northern Hemisphere and Southern Hemisphere to be preferentially oriented either towards or away from the Sun, depending on the time of year.

Check it out for yourself with this animation:

The consequence of all of this, is that amount of short wave radiation received from the Sun at the top of the Earth's atmosphere varies as a function of both time of day and season:

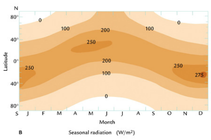


Figure 1.9: Seasonal Distribution of Net Solar Radiation Received at Earth's Surface With Latitude
Credit: Ruddiman, *Earth's Climate: Past and Future* (W.H. Freeman, 2001)

Subtle changes in the Earth's orbital geometry (i.e., changes in the tilt of the axis, the degree of ellipticality of the orbit, and the slow precession of the orbit) are responsible for the coming and going of the ice ages over tens of thousands of years. We will revisit this topic later in the course.

Overview of the Climate System (part 3)

Atmospheric Circulation

We have seen above that the distribution of solar insolation over the Earth's surface changes over the course of the seasons, with the Sun, in a relative sense, migrating south and then north of the equator over the course of the year—that annual migration, between 23S and 23N, defines the region we call *the tropics*. As the heating by the Sun migrates south and north within the tropics over the course of the year, so does the tendency for rising atmospheric motion. As we have seen, warmer air is less dense than cold air, and where the Sun is heating the surface there is a tendency for *convective instability*, i.e., the unstable situation of having relatively light air underlying relatively heavy air. Where that instability exists, there is a tendency for rising motion in the atmosphere, as the warm air seeks to rise above the colder air. As a result, there is a tendency for rising air (and with it, rainfall) in a zone of low surface pressure known as the *Intertropical Convergence Zone* or *ITCZ*, which is centered roughly at the equator, but shifts north and south with the migration of the Sun about the equator over the course of the year. Due to the greater thermal inertia of the oceans relative to the land surface, the response to the shifting solar heating is more sluggish over the ocean, and the ITCZ shows less of a latitudinal shift with the seasons. By contrast, over the largest land masses (e.g., Asia), the seasonal shifts can be quite pronounced, resulting in dramatics shifts in wind and rainfall patterns such as the *Indian monsoon*.

The air rising in the tropics then sinks in the subtropics, forming a subtropical band of high surface pressure and low precipitation associated with the prevailing *belt of deserts* in the subtropics of both hemispheres. The resulting pattern of circulation of the atmosphere is known as the *Hadley Cell circulation*. In sub-polar latitudes, there is another region of low surface pressure, associated again with rising atmospheric motion and rainfall.

This region is known as the *polar front*. These belts of high and low atmospheric surface pressure, and the associated patterns of atmospheric circulation also shift south and north over the course of the year in response to the heating by the Sun. You can explore the atmospheric patterns using this map:

We have seen above that there is an imbalance between the absorbed incoming *short wave* solar radiation and the emitted outgoing *long wave* terrestrial radiation, with a relative surplus within the tropics and a relative deficit near the poles. We, furthermore, noted that the atmosphere and ocean somehow relieve this imbalance by transporting heat laterally, through a process known as *heat advection*. We are now going to look more closely at how the atmosphere accomplishes this transport of heat. We have already seen one important ingredient, namely the Hadley Cell circulation, which has the net effect of transporting heat poleward from where there is a surplus to where there is a deficit.

Wind patterns in the extratropics also serve to transport heat poleward. The lateral wind patterns are primarily governed by a balance between the previously discussed pressure gradient force (acting in this case laterally rather than vertically), and the *Coriolis force*, an effective force that exists due to the fact that the Earth is itself rotating. This balance is known as the *geostrophic* balance.

The Coriolis force acts at right angles to the direction of motion: 90 degrees to the right in the Northern Hemisphere and 90 degrees to the left in the Southern Hemisphere. The pressure gradient force is directed from regions of high surface pressure to regions of low surface pressure. As a consequence, geostrophic balance leads to winds in the mid-latitudes, between the subtropical high pressure belt and the sub-polar low pressure belt of the polar front, blowing from west to east. We call these *westerly* winds. For reasons that have to do with the vertical thermal structure of the atmosphere, and the combined effect of the geostrophic horizontal force balance and hydrostatic vertical force balance in the atmosphere, the westerly winds become stronger aloft, leading to the intense regions of high wind known as the *jet streams* in the mid-latitude upper troposphere.

Conversely, winds in the tropics tend to blow from east to west. These are known as *easterly winds* or, by the perhaps more familiar term, the *trade winds*. In the Northern Hemisphere, geostrophic balance implies counter-clockwise rotation of winds about low pressure centers and clockwise rotation of winds about high pressure centers. The directions are opposite in the Southern Hemisphere.

Due to the effect of friction at the Earth's surface, there is an additional component to the winds which blows *out* from high pressure centers and *in* towards low pressure centers. The result is spiraling in (convergence) towards low pressure centers and a spiraling out (divergence) about high pressure centers. The convergence of the winds toward the low pressure centers is associated with the rising atmospheric motion that occurs within regions of low surface pressure. The divergence of the winds away from the high pressure centers is associated with the sinking atmospheric motion that occurs within regions of high atmospheric pressure.

The inward spiraling low pressure systems in mid-latitudes constitute the polar front, which separates the coldest air masses near the poles from the warmer air masses in the subtropics. In fact, it is the unstable property of having clashing air masses with vastly different temperature characteristics, known as *baroclinic instability*, that is responsible for the existence of extratropical cyclones. The energy that drives the extratropical cyclones comes from the work done as surface air is lifted along frontal (i.e., cold front and warm front) boundaries. These extratropical storm systems relieve high-latitude deficit of radiation by mixing cold polar and warm subtropical air and, in so doing, transporting heat poleward, along the latitudinal temperature gradient.

You can explore the resulting large-scale pattern of circulation of the global atmosphere in this animation:

Ocean Circulation (Gyres, Thermohaline Circulation)

While we have focused primarily on the atmosphere thus far, the oceans, too, play a key role in relieving the radiation imbalance by transporting heat from lower to higher latitudes. The oceans also play a key role in both climate variability and climate change, as we will see. There are two primary components of the ocean circulation. The first component is the horizontal circulation, characterized by *wind-driven ocean gyres*.

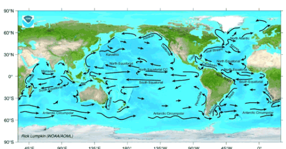


Figure 1.10: Global Ocean Currents.
Credit: NOAA [13]

The major surface currents are associated with the ocean gyres. These include the warm poleward *western boundary* currents such as the *Gulf Stream*, which is associated with the *North Atlantic Gyre*, and the *Kuroshio Current* associated with the *North Pacific Gyre*. These gyres also contain cold equatorward *eastern boundary currents* such as the *Canary Current* in the eastern North Atlantic and the *California Current* in the western North Atlantic. Similar current systems are found in the Southern Hemisphere. The horizontal patterns of ocean circulation are driven by the alternating patterns of wind as a function of latitude, and, in particular, by the tendency for westerly winds in mid-latitudes and easterly winds in the tropics, discussed above.

An important additional mode of ocean circulation is the *thermohaline circulation*, which is sometimes referred to as the *meridional overturning circulation* or *MOC*. The circulation pattern is shown below:

circulation or MOC. The circulation pattern is shown below.

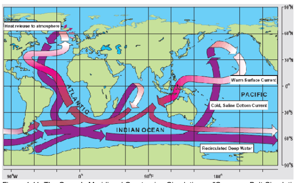


Figure 1.11: The Ocean's Meridional Overturning Circulation or "Conveyor Belt Circulation"
Credit: Image courtesy of Windows to the Universe [14]

By contrast with the horizontal gyre circulations, the MOC can be viewed as a vertical circulation pattern associated with a tendency for sinking motion in the high-latitudes of the North Atlantic, and rising motion more broadly in the tropics and subtropics of the Indian and Pacific ocean. This circulation pattern is driven by contrasts in density, which are, in turn, largely due to variations in both temperature and salinity (hence the term *thermohaline*). The sinking motion is associated with relatively cold, salty surface waters of the sub-polar North Atlantic, and the rising motion with the relatively warm waters in the tropical and subtropical Pacific and Indian ocean.

The picture presented in Figure 1.11 above is a highly schematized and simplistic description of the actual vertical patterns of circulation in the ocean. Nonetheless, the *conveyor belt* is a useful mnemonic. The northward surface branch of this circulation pattern in the North Atlantic is sometimes erroneously called the Gulf Stream. The Gulf Stream, as discussed above, is part of the circulating waters of the wind-driven ocean gyre circulation. By contrast, the northward extension of the thermohaline circulation in the North Atlantic, is rightfully referred to as the *North Atlantic Drift*. This current system represents a net transport of warm surface waters to higher latitudes in the North Atlantic and is also an important means by which the climate system transports heat poleward from lower latitudes. Changes in this current system are speculated as having played a key role in past and potential future climate changes, as will be explored later in this course.

Other Fundamental Principles

Reading Assignment

Before we go any further, please make sure that you have read the following document:

- From the IPCC Fifth Assessment Working Group 1 [2], read the *Introduction* (p.4), section B.5, *Observed Changes in the Climate System, Carbon and Other Biogeochemical Cycles* (p. 11-12), and section C, *Drivers of Climate Change* (p. 13-14) of the *Summary for Policy Makers* [3].

As you read, pay particular note to the discussion of natural vs. human drivers of climate, and the historic changes in the natural and human factors influencing climate. Consider the following questions:

Is modern climate change driven solely by humans?

Can we explain global warming in terms of natural factors?

Are these questions reasonable, or are they too simplistic?

Natural vs. Human Forcing

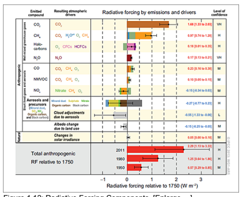


Figure 1.12: Radiative Forcing Components. [Enlarge [15]]
Credit: IPCC, 2013

Let us consider more closely the above figure (Figure 1.12) from the IPCC Summary for Policy Makers . There is a lot of information packed in this figure. Figure 1.12 summarizes the relative impacts of various natural and human forcing factors on the Earth's climate. Later on in the course, we will look at how these forcing factors are likely to have impacted global mean temperature trends over the past century. In the meantime, we can make some rough assessment of the relative importance of the different factors as gauged by their estimated *radiative forcing* — measured by the energy per unit time that a given forcing factor exerts per square meter of the Earth's surface. The first thing to pay attention to is whether the indicated forcing factor is a warming factor (black dot to the right of zero) or cooling factor (black dot to the left of zero). The next thing to take note of is how high or low the forcing associated with that factor is, as indicated by the length of the bar. Finally, take note of the error bars (these are shown as the horizontal "barbell" symbols) indicating whether the factor in question is relatively well known, or relatively uncertain.

The forcings are separated into two fundamentally different categories: *anthropogenic* (that is, human-caused) and *natural*. You may be surprised to learn that while greenhouse gases are the primary anthropogenic forcing, there are other notable anthropogenic forcing contributions. Indeed, if one computes the net effect of anthropogenic aerosols (primarily sulphate) produced by industrial activity, adding together the *direct* and *indirect* effects of these aerosols, the total *negative* global radiative forcing (roughly -0.8 W/m^2) is nearly half as large as the positive radiative forcing (roughly 1.7 W/m^2) due to human-caused CO_2 concentration increases (though the uncertainty associated with the most recent estimate of aerosol forcing is quite large). As we will see in later lectures, the cooling effect of these aerosols offset a substantial fraction of anthropogenic greenhouse warming over the past century.

One important historical natural forcing of climate is *not* shown in this diagram. This is the cooling effect of volcanic eruptions due to reflective aerosol injected into the stratosphere. Unlike other forcings, this volcanic forcing is episodic, rather than continuous in nature. Explosive volcanic eruptions may have a cooling effect on climate for several years. If there is a large number of eruptions over a sustained period of time, this can have an overall cooling impact on climate. We will revisit this issue in Lesson 4.

Feedback Mechanisms

The response of the climate to forcing, whether natural or human-caused, would be far more modest than it is, were it not for the influence of *feedback mechanisms*. Feedback mechanisms are mechanisms within the climate system that act to either attenuate (negative feedback) or amplify (positive feedback) the response to a given forcing. On balance, the feedbacks are believed to be positive in the sense that the response of the climate system to a positive forcing is greater than one would expect from the forcing alone, because of the net warming effect arising from these responses.

The principle feedback mechanisms relevant to climate change on historical timescales, are:

1. The *water vapor feedback*. Warming atmosphere can hold larger amounts of water vapor. Since water vapor is a greenhouse gas, this leads to further warming:a **Positive Feedback** [16]
2. The *ice-albedo feedback*. Surface of the Earth has less snow/ice as it warms, leading to less reflection and greater absorption of incoming solar radiation: a **Positive Feedback**
3. The *cloud radiative feedbacks*. There are different competing effects:
   1. Warmer atmosphere produces more *low* clouds. The primary impact of more low clouds would be to reflect more solar radiation out to space: a **Negative Feedback** [17]
   2. Warmer atmosphere produces more *high* clouds, like cirrus. The primary impact of such thin, high clouds is to increase the greenhouse effect due to their ability to trap much of the outgoing longwave terrestrial radiation while remaining largely transparent to incoming shortwave solar radiation: a **Positive Feedback**

On balance, it has been believed that the negative cloud radiative feedbacks win over the positive cloud radiative feedbacks - though the low cloud feedbacks are quite uncertain, and the overall cloud radiative feedback could very well be positive.

The net effect of all these feedbacks is *positive* and serves to increase the warming due a particular external forcing (be it increased greenhouse gas concentrations due to fossil fuel emissions, increased solar output, or some other external forcing) beyond what would be expected purely from that factor alone. For example, a doubling of CO_2 concentrations relative to pre-industrial levels would, in the absence of feedbacks, lead to roughly 1.25°C warming. However, our best estimates indicate that the positive water vapor feedback would add about 2.5°C additional warming, while the positive ice albedo feedback adds about 0.6°C warming. While substantially more uncertain, the negative cloud radiative feedback could lead to just under 2°C cooling. Add up the numbers, and the total comes to about 2.5°C warming (actually, current generation climate models average closer to 3°C).

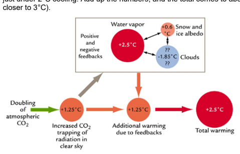


Figure 1.13: Climate Feedback Mechanisms
Credit: Ruddiman, *Earth's Climate: Past and Future* (W.H. Freeman, 2001)

This quantity--how much we expect the Earth to warm once it equilibrates to a doubling of greenhouse gas concentrations--is known as the *equilibrium climate sensitivity*. We will explore this key concept in more detail in subsequent lectures.

The Carbon Cycle

The traditional concept of climate sensitivity envisions the concentration of CO_2 and other greenhouse gases as specified (i.e., doubled from some initial level), and calculates the expected warming. This construction is somewhat artificial, however, because activities, such as fossil fuel burning, do not directly regulate the *concentration* of CO_2 or other greenhouse gases, but instead govern the atmospheric *emissions*, which can interact with the climate system. For example, life on land and in the ocean can both take up and give off CO_2: CO_2 is taken up during *photosynthesis* — production of organic matter by green plants, and given off during *respiration* (or *remineralization*) — a reverse process during which organic matter is decomposed. As climate becomes warmer, the living organisms are affected by the change, e.g., green plants might consume more CO_2 because the growing season becomes longer and the plants have more time for photosynthesis; or, on the other hand, warmer temperatures might induce bacterial activity and the rates of decay of organic matter, causing an increase in the CO_2 emissions. In general, the Earth system processes of chemical, physical, or biological origin that emit CO_2 to the atmosphere are referred to as carbon *sources,* while those that take up CO_2 from the atmosphere are referred to as carbon *sinks* or *losses*. Climate change affects the characteristics of living things, as well as other components of the climate system, such as, for example, the overturning ocean circulation (which helps to sequester atmospheric carbon in the deep ocean), and therefore can influence various carbon sources and sinks that exist within the Earth system.

Figure 1.14: The Carbon Cycle.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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We refer to the amount of emitted CO_2 that actually stays in the atmosphere as the *airborne* fraction of CO_2. So far, only roughly half of our carbon emissions remain airborne. The other half has been absorbed by carbon sinks. The primary carbon *sink* is the upper ocean, which has absorbed roughly 25-30% of the CO_2, while the terrestrial biosphere has absorbed another 15-20% of the CO_2.

These *sinks* are not constant over time, however. Numerous studies indicate that both the upper ocean and terrestrial biosphere are likely to become less able to absorb and hold additional CO_2 as the globe warms. Were this to happen, the airborne fraction of CO_2 in the atmosphere would increase, and CO_2 would accumulate in the atmosphere more quickly for a given rate of emissions. Such responses are known as the *carbon cycle feedbacks*, because they have the ability to influence the accumulation of CO_2 in the atmosphere.

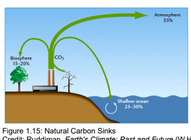


Figure 1.15: Natural Carbon Sinks
Credit: Ruddiman, *Earth's Climate: Past and Future* (W.H. Freeman, 2001)

The existence of carbon cycle feedbacks forces us to reconsider the concept of *climate sensitivity* discussed earlier. Consider for example the accumulated carbon emissions that we might calculate would lead to a doubling of CO_2 in the atmosphere *in the absence* of carbon cycle feedbacks. As the climate warms, the positive carbon cycle feedbacks discussed above would cause the airborne fraction to increase. As a result, the final increase in atmospheric CO_2 would be greater than the originally calculated doubling. Accordingly, there would be even more warming than one would estimate from applying the standard concept of equilibrium climate sensitivity to the original estimated slug of carbon emissions.

Such complications lead to the more general notion of the *Earth System sensitivity*. We will revisit these concepts later in the course.

Lesson 1 Discussion

Activity

Directions

At this point, you have completed the Course Orientation and the first lesson for METEO 469. Let's talk! Please participate in an online discussion of the course in general and of the material we have covered thus far. Please share your thoughts about the general topic of this course and what you hope to learn. Also, please discuss the material presented in Lesson 1.

This discussion will take place in a threaded discussion forum in Canvas (see the Canvas Guides [18] for the specific information on how to use this tool) over approximately a week-long period of time. Since the class participants will be posting to the discussion forum at various points in time during the week, you will need to check the forum frequently in order to fully participate. You can also subscribe to the discussion and receive e-mail alerts each time there is a new post.

Please realize that a discussion is a group effort and make sure to participate early in order to give your classmates enough time to respond to your posts.

Post your comments addressing some aspect of the material that is of interest to you and respond to other postings by asking for clarification, asking a follow-up question, expanding on what has already been said, etc. For each new topic you are posting, please try to start a new discussion thread with a descriptive title, in order to make the conversation easier to follow.

Suggested topics

The purpose of the discussion is to facilitate a free exchange of thoughts and opinions among the students, and you are encouraged to discuss any topic within the general discussion theme that is of interest to you. If you find it helpful, you may also use the topics suggested below.

General discussion of METEO 469

- Why are you taking this course?
- Why are you interested in climate change?
- Is it important to you to understand the scientific basis of climate science?
- Is it important to you to learn about impacts of climate change?
- What do you hope to learn in this course?

Lesson 1: Introduction to Climate and Climate Change

- In your understanding, what is the difference between the climate and the weather? Why is it important to differentiate between the two?
- Discuss natural v.s. anthropogenic climate change.
- What are feedback mechanisms in regards to the climate system? Can we know all feedback mechanisms in our climate system? Which mechanisms are considered most important and why?
- Discuss the concept of climate sensitivity. Why is this concept useful?
- Discuss the components of the carbon cycle and how carbon sources and sinks influence atmospheric CO_2 levels.

Submitting your work

1. Go to Canvas.
2. Go to the *Home* tab.
3. Click on *Lesson 1 discussion: General Discussion of METEO 469*.
4. Post your comments and responses.

Grading criteria

You will be graded on the quality of your participation. See the online discussion grading rubric [19] for the specifics on how this assignment will be graded. Please note that you will not receive a passing grade on this assignment if you wait until the last day of the discussion to make your first post.

Lesson 1 Summary

We have reviewed the essential basic concepts necessary for understanding climate change and global warming, including:

- the concepts of climate, climate change, and global warming;
- the components of the Earth's climate system: the atmosphere, oceans, cryosphere, and biosphere;
- the structure and composition of the Earth's atmosphere;
- the concepts of radiation and energy balance;
- the nature of the circulation of the atmosphere and the oceans;
- the concept of radiative forcing;
- climate feedbacks and climate sensitivity;
- carbon cycle feedbacks and the Earth System sensitivity.

We are now well equipped to begin digging into the details. Our first foray will be into the world of *climate observations*. What data are available that can inform our understanding of how climate has changed over historic time? What indirect data are available that place historical observation in a longer-term context? How do we analyze such data to assess whether there is indeed "climate change"? This will be our next topic.

Reminder - Complete all of the module tasks!

You have finished Lesson 1. Double-check the list of requirements on the first page of this lesson to make sure you have completed all of the activities listed there before beginning the next lesson.

Lesson 2 - Climate Observations, part 1

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

About Lesson 2

How do we know that climate change is taking place? Or that the factors we believe to be driving climate change, such as greenhouse gas concentrations, are themselves changing?

To address these questions, we turn first to instrumental measurements documenting changes in the properties of our atmosphere over time. These measurements are not without their uncertainties, particularly in earlier times. But they can help us to assess whether there appear to be trends in measures of climate and the factors governing climate, and whether the trends are consistent with our expectations of what the response of the climate system to human impacts ought to look like.

What will we learn in Lesson 2?

By the end of Lesson 2, you should be able to:

- discuss the various modern observational data characterizing changes in surface and atmospheric temperature over the historical period;
- discuss the nature of the uncertainties in the observational record of past climate; and
- perform simple statistical analyses to characterize trends in, and relationships between, data series.

What will be due for Lesson 2?

Please refer to the Syllabus for the specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 2. Detailed directions and submission instructions are located within this lesson.

- Read:
  - IPCC Fifth Assessment Report, Working Group 1 [2]
  - Summary for Policy Makers [3]
    - B Observed Changes in the Climate System: p. 4
    - B.1 Atmosphere: p. 5-8
  - Dire Predictions, v.2: p. 34-35, 38-39, 80-81
- Problem Set #1: Perform basic statistical analyses of climate data.

Questions?

If you have any questions, please post them to our *Questions?* discussion forum (not e-mail), located under the Home tab in Canvas. The instructor will check that discussion forum daily to respond. While you are there, feel free to post your own responses if you can help with any of the posted questions.

Observed Changes in Greenhouse Gases

Before we assess the climate data documenting changes in the climate system, we ought to address the question — is there evidence that greenhouse gases, purportedly responsible for observed warming, are actually changing in the first place?

Thanks to two legendary atmospheric scientists, we know that there is such evidence. The first of these scientists was Roger Revelle [20].

Roger Revelle.
Credit: Cambridge Forum Speakers [21]

Revelle, as we will later see, made fundamental contributions to understanding climate change throughout his career. Less known, but equally important, was the tutelage and mentorship that Revelle provided to other climate researchers. While at the Scripps Institution for Oceanography at the University of California in San Diego, Revelle encouraged his colleague Charles David Keeling [22] to make direct measurements of atmospheric CO_2 levels from an appropriately selected site.

Think About It!

Why do you suppose it is adequate to make measurements of atmospheric CO_2 from a single location as an indication of changing global concentrations?

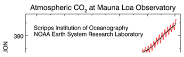
Click for answer.

Revelle and Keeling settled on the top of the mountain peak Mauna Loa on the big island of Hawaii, establishing during the *International Geophysical Year* of 1958 an observatory that would be maintained by Keeling and his crew for the ensuing decades.

Charles David Keeling.
Credit: NOAA [23]

Mauna Loa Observatory.
Credit: NOAA [24]

From this location, Keeling and crew would make continuous measurements of atmospheric CO_2 from 1958 henceforth. Since then, long-term records have been established in other locations over the globe as well. The result of Keeling's labors is arguably the most famous curve in all of atmospheric science, the so-called *Keeling Curve*. That curve shows a steady increase in atmospheric CO_2 concentrations from about 315 ppm when the measurements began in 1958 to about 400 today (and climbing by about 2 ppm per year presently).



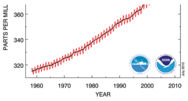


Figure 2.1: Atmospheric CO_2 at Mauna Loa Observatory
Credit: NOAA [25]

You might be wondering at this point, how do we know that the increase in CO_2 is not natural? For one thing, as you already have encountered in your readings, the recent levels are unprecedented over many millennia. Indeed, when we cover the topic of *paleoclimate*, we will see that the modern levels are unprecedented over many *hundreds of thousands* of years, and probably several million years. We will also see that there is a long-term relationship between CO_2 and temperature, though the story is not as simple as you might think.

But there is other more direct evidence that the source of the increasing CO_2 is indeed human, i.e., anthropogenic. It turns out that carbon that gets buried in the earth from dying organic matter and eventually turns into fossil fuels, tends to be *isotopically light*. That is, nature has a preference for burying carbon that is depleted of the heavier, ^{13}C, carbon isotope. Fossil fuels are thus relatively rich in the lighter isotope, ^{12}C. However, natural atmospheric CO_2 produced by respiration (be it animals like us, or plants which both respire and photosynthesize) tends to have a greater abundance of the heavier ^{13}C isotope of carbon. If the CO_2 increase were from natural sources, we would therefore expect the ratio of ^{13}C to ^{12}C to be getting higher. But instead, the ratio of ^{13}C to ^{12}C is getting lower as CO_2 is building up in our atmosphere -- i.e., the ratio bears the *fingerprint* of anthropogenic fossil fuel burning.

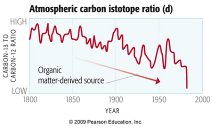


Figure 2.2: Graph of carbon-13 to carbon-12 Ratio from 1800 - 2000.
Credit: Mann and Kump, *Dire Predictions: Understanding Global Warming* (DK, 2008, 2015)

Of course, CO_2 is not the only greenhouse gas whose concentrations are rising due to human activity. A combination of agriculture (e.g., rice cultivation), livestock raising, and dam construction led to substantial increases in methane (CH_4) concentrations. Agricultural practices have also increased the concentration of nitrous oxide (N_2O).

Using air bubbles in ice cores, we can examine small bits of atmosphere trapped in ice, as it accumulated back in time, to reconstruct the composition of the ancient atmosphere, including the past concentrations of greenhouse gases. The ice core evidence shows that the rise over the past two centuries in the concentrations of the greenhouse gases mentioned above is unprecedented for at least the past 10,000 years. Longer-term evidence suggests that concentrations are higher now than they have been for hundreds of thousands of years, and perhaps several million years.

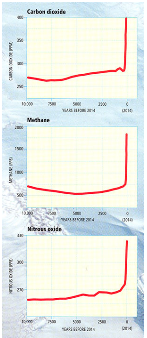


Figure 2.3: Changes in Greenhouse Gases Record in Ice Cores.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Reading Assignment

Before we go any further, please read the following document:

- From the IPCC Fifth Assessment Report, Working Group 1 [2], read section B, Observed Changes in the Climate System (p. 4), and section B.1, Atmosphere (p. 5-8) of the Summary for Policy Makers [3].

As you read, please pay particular attention to:

- *the variety of sources of data,*
- *internal consistency among the various data streams with regard to our changing climate.*

Modern Surface Temperature Trends

Instrumental surface temperature measurements consisting of thermometer records from land-based stations, islands, and ship-board measurements of ocean surface temperatures provide us with more than a century of reasonably global estimates of surface temperature change. Some regions, like the Arctic and Antarctic, and large parts of South America, Africa, and Eurasia, were not very well sampled in earlier decades, but records in these regions become available as we move into the mid and late 20th century.

Temperature variations are typically measured in terms of *anomalies* relative to some *base period*. The animation below is taken from the NASA Goddard Institute for Space Studies [26] in New York (which happens to sit just above "Tom's Diner" [27] of *Seinfeld* fame), one of several scientific institutions that monitor global temperature changes. It portrays how temperatures around the globe have changed in various regions since the late 19th century. The temperature data have been averaged into 5 year blocks, and reflect variations relative to a 1950-1980 base period, i.e., *warm* regions are warmer than the 1950-1980 average, while *cold* regions are colder than the 1950-1980 average, by the magnitudes shown. You may note a number that appears in the upper right corner of the plot. That number indicates the average temperature anomaly over the entire globe at any given time, again, relative to the 1950-1980 average.

Take some time to explore the animation on your own. You may want to go through it several times so you can start to get a sense of just how rich and complex the patterns of surface temperature variations are. Do you see periodic intervals of warming and cooling in the eastern equatorial Pacific? What might that be? [We will talk about the phenomenon in upcoming lessons].

Take note of any particularly interesting patterns in space and time that you see as you review the animation. You can turn your sound off the first few times so you do not hear the annotation of the animation. Then, when you are ready, turn the sound on and you can hear Michael Mann's take.

Credit: NASA's Goddard Institute for Space Studies [28]

We can average over the entire globe for any given year and get a single number, the global average temperature. Here is the curve we get if we plot out that quantity. Note that in the plot below the average temperature over the base period has been added to the anomalies, so that the estimate reflects the surface temperature of the Earth itself.

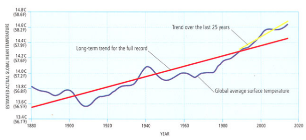

Figure 2.4: Trends in Global Average Surface Temp. 1860-2015.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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We can see that the Earth has warmed a little less than 1°C (about 1.5 F) since widespread records became available in the mid-19th century. That this warming has taken place is essentially incontrovertible from a scientific point of view. What is the *cause* of this warming? That is a more difficult question, which we will address later.

We discussed above the cooling that is evident in parts of the Northern Hemisphere (particularly over the land masses) from the 1940s-1970s. There was a time during the mid 1970s, when *some* scientists thought the globe might be entering into a long-term *cooling trend*. There was a reason to believe that might be the case. In the absence of other factors, changes in the Earth's orbital geometry did favor the descent (albeit a very slow one!) into the next ice age. Also, the buildup of atmospheric aerosols, which, as we will explore, can have a large regional cooling impact, favored cooling. Precisely how these cooling effects would balance out against the warming impact of greenhouse gases was not known at the time.

Some critics claim that if the scientific community thought were were entering into another Ice Age in the 1970s, why should we trust the scientists now about global warming? In fact, it was far from a scientific consensus [29] within the scientific community in the mid 1970s that we were headed into another Ice Age. Some scientists speculated this was possible, but the prevailing viewpoint was that increasing greenhouse gas concentrations and warming would likely win out.

We know that, indeed, the short term cooling trend for the Northern Hemisphere continents ended in the 1970s, and, since then, global warming has dominated over any cooling effects.

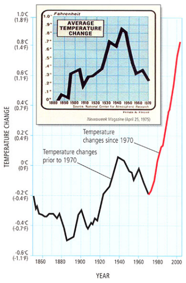

Figure 2.5: Northern Hemisphere Continental Temperature Trends.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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As mentioned earlier, we cannot deduce the *cause* of the observed warming solely from the fact that the globe is warming. However, we can look for possible clues. Just like forensics experts, climate scientists refer to these clues as *fingerprints*. It turns out that natural sources of warming give rise to different patterns of temperature change than human sources, such as increasing greenhouse gases. This is particularly true when we look at the *vertical pattern* of warming in the atmosphere. This is our next topic.

Vertical Temperature Trends

As alluded to previously, the vertical pattern of observed atmospheric temperature trends provides some important clues in establishing the underlying cause of the warming. While upper air temperature estimates (from weather balloons and satellite measurements) are only available for the latter half of the past century, they reveal a remarkable pattern. The lower part of the atmosphere — the *troposphere,* has been warming along with the surface. However, once we get into the *stratosphere*, the temperatures have actually been decreasing! As we will learn later when we focus on the problem of climate signal *fingerprinting*, certain forcings are consistent with such a vertical pattern of temperature changes, while other forcings are not.

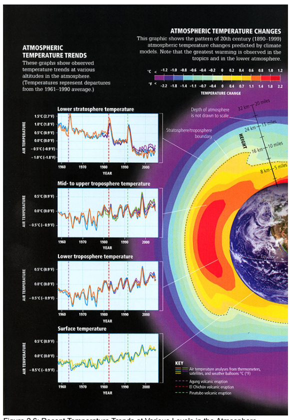

Figure 2.6: Recent Temperature Trends at Various Levels in the Atmosphere.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Think About It!

Care to venture a guess as to which forcing might be most consistent with this vertical pattern of temperature change?

Click for answer.

Historical Variations in Precipitation and Drought

Recall our discussion of the general circulation of the atmosphere [30] from Lesson #1.

There we learned that the circulation of the atmosphere is driven by the contrast in surface heating between the equator and the poles. That contrast results from the difference between incoming short wave solar heating and outgoing loss from the surface through various modes of energy transport including radiational heat loss as well as heat loss through convection and latent heat release through evaporation.

It, therefore, stands to reason that climate change — which in principle involves changing the balance between incoming and outgoing radiative loss via changes in the greenhouse effect — is likely to alter the circulation of the atmosphere itself, and thus, large-scale precipitation patterns. The observed changes in precipitation patterns are far *noisier* (very variable and difficult to interpret) than temperature changes, however. Regional effects related to topography (e.g., mountain ranges that force air upward leading to wet windward and dry leeward conditions), ocean-atmosphere heating contrasts that drive regional circulation patterns, such as monsoons, etc., lead to very heterogeneous patterns of changes in rainfall, in comparison with the pattern of surface temperature changes.

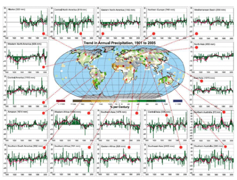


Figure 2.7: Trends in Annual Precipitation, 1901-2005 [Enlarge [31]].
Credit: IPCC Fourth Assessment Report, Chapter 3, Figure 3.14

We might expect certain reasonably simple patterns to emerge, nonetheless. As we shall see in a later lesson [32] looking at climate change projections, climate models predict that atmospheric circulation cells and storm tracks migrate poleward, shifting patterns of rainfall between the equator and poles. The subtropics and middle latitudes tend to get dryer, while the sub-polar latitudes get wetter (primarily in winter). The equatorial region actually is predicted to get wetter, simply because the rising motion that occurs there squeezes out more rainfall from the warmer, moister lower atmosphere. If we average the observed precipitation changes in terms of trends in different latitudinal bands, we can see some evidence of the changes.

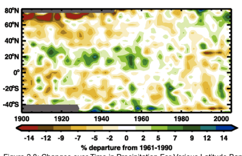


Figure 2.8: Changes over Time in Precipitation For Various Latitude Bands
Credit: IPCC Fourth Assessment Report, Chapter 3, Figure 3.15

For example, we see that over time the high northern latitudes (60-80N) are getting wetter, while the subtropical and middle latitudes of the Northern Hemisphere are getting dryer. However, there is a lot of variability from year to year, and from decade to decade, making it difficult to clearly discern whether the theoretically predicted changes are yet evident.

Drought, as we will see, does not simply follow rainfall changes. Rather, it reflects a combination of both rainfall and temperature influences. Decreased rainfall can lead to warmer ground temperatures, increased evaporation from the surface, decreased soil moisture, and thus drying.

Like rainfall, regional patterns of drought are complicated and influenced by a number of different factors. However, the combination of shifting rainfall patterns and increased evaporation has led to very pronounced increases in drought in subtropical regions, and even in many tropical and sub-polar regions, where rainfall shows little trend (as indicated in the earlier graphic) but warmer temperatures have led to decreased soil moisture. These broad trends are seen in measurements of the *Palmer Drought Severity Index* -- an index that combines the effects of changing rainfall and temperature to estimate soil moisture content; the more negative the index, the stronger the drought.

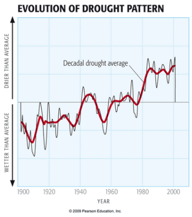


Figure 2.9: Evolution of Drought Pattern.
Credit: Pearson, 2009

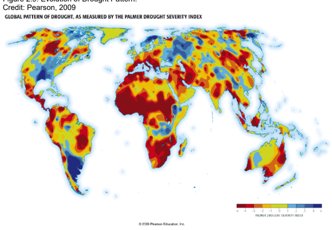


Figure 2.10: Global Pattern of Drought, as Measured by the Palmer Drought Severity Index.
Credit: Pearson, 2009

In the next lesson, we will assess evidence for changes in extreme weather events, such as heat waves, floods, tropical cyclone activity, etc. In the meantime, however, we are going to digress a bit and discuss the topic of how to analyze data for inferences into such matters as discerning whether or not trends are evident in particular data sets, and whether it is possible to establish a relationship between two or more different data sets.

Review of Basic Statistical Analysis Methods for Analyzing Data - Part 1

Now that we have looked at the basic data, we need to talk about how to *analyze* the data to make inferences about what they may tell us.

The sorts of questions we might want to answer are:

- Do the data indicate a *trend*?
- Is there an apparent *relationship* between two or more different data sets?

These sorts of questions may seem simple, but they are not. They require us, first of all, to introduce the concept of *hypothesis testing*.

To ask questions of a data set, one has to first formalize the question in a meaningful way. For example, if we want to know whether or not a data series, such as global average temperatures, display a trend, we need to think carefully about what it means to say that a data series has a trend!

This leads us to consider the concept of the *null hypothesis*. The null hypothesis states what we would expect purely from chance alone, in the absence of anything interesting (such as a trend) in the data. In many circumstances, the null hypothesis is that the data are the product of being randomly drawn from a *normal distribution*, what is often called a *bell curve*, or sometimes, a *Gaussian distribution* (after the great mathematician Carl Friedrich Gauss [33]):

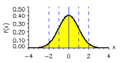


Figure 2.11: Gaussian Distribution.

In the normal distribution shown above, the average or *mean* of the data set has been set to zero (that is where the peak is centered), and the *standard deviation* (*s.d.*), a measure of the typical amplitude of the fluctuations, is set to one. If we draw random samples from such a distribution, then roughly 68% of the time the values will fall within 1 s.d. of the mean (in the above example, that is the range -1 to +1). That means that roughly 16% of the time the data will fall above 1 s.d., and roughly 16% of the time the data will fall below 1 s.d. About 95% of the time, the randomly drawn values will fall within 2 s.d. (i.e., the range -2 to +2 in the above example). That means only 2.5% of the time the data will fall above 2 s.d. and only 2.5% of the time below 2 s.d. For this reason, the 2 s.d. (or *2 sigma*) range, is often used to characterize the region we are relatively confident the data should fall in, and the data that fall outside that range are candidates for potentially interesting anomalies.

Random Time Series

Here is an example of what a random data series of length *N* = 200 which we will call ε(*t*), drawn from a simple normal distribution with mean zero and standard deviation one looks like (for example, you can think of this data set as a 200 year long temperature anomaly record).

$$Y_t = \varepsilon(t)$$

(1)

This sort of noise is called *white noise* because there is no particular preference for either higher-frequency or lower-frequency fluctuations. The fluctuations have equal amplitude.

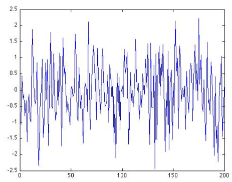


Figure 2.12(1). *N*=200 years of Gaussian White Noise.

There is another form of random noise, known as *red noise* because the long-term fluctuations have a greater relative magnitude than short-term fluctuations (just as red light is dominated by low-frequency visible wavelengths of light).

A simple model for Gaussian red noise takes the form

$$Y_t = \rho \cdot Y_{t-1} + \varepsilon(t)$$

(2)

where ε(*t*) is Gaussian white noise. As you can see, a red noise process tends to integrate the white noise over time. It is this process of integration that leads to more long-term variation than would be expected for a pure white noise series. Visually, we can see that the variations from one year to the next are not nearly as erratic. This means that the data have fewer *degrees of freedom (N′)* than there are actual data points (*N*). In fact, there is a simple formula relating *N′* and *N*:

$$N' = N\frac{1-\rho}{1+\rho}$$

(3)

The factor $(1-\rho)/(1+\rho)$ measures the "redness" of the noise. Let us consider again a random sequence of length *N* = 200 but this time it is "red" with the value ρ = 0.6. The same random white noise sequence used previously is used in equation 2 for ε(*t*):

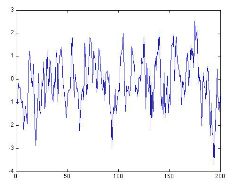


Figure 2.12(2): *N*=200 years of Gaussian 'red noise' with ρ=0.6

Self-Check

How many distinct peaks and troughs can you see in the series now?

Click for answer.

Self-Check

How many degrees of freedom *N* ' are there in this series?

Click for answer.

As ρ gets larger and larger, and approaches one, the low-frequency fluctuations become larger and larger. In the limit where ρ = 1, we have what is known as a *random walk* or *Brownian motion*. Equation 2 in this case becomes just:

$$y_t = y_{t-1} + \varepsilon(t)$$

(4)

You might notice a problem when using equation 3 in this case. For ρ = 1, we have *N′* = 0! There are no longer any effective degrees of freedom in the time series. That might seem nonsensical. But there are other attributes that make this a rather odd case as well. The time series, it turns out, the time series. That might seem nonsensical. But there are other attributes that make this a rather odd case as well. The time series, it turns out, now has an infinite *standard deviation*!

Let's look at what our original time series looks like when we now use ρ = 1:

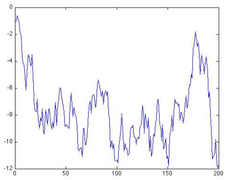


Figure 2.12(3): *N*=200 years of Gaussian 'red noise' with ρ=1, i.e., a 'random walk'.

As you can see, the series starts out in the same place, but immediately begins making increasingly large amplitude long-term excursions up and down. It might look as if the series wants to stay negative. But if we were to continue the series further, it would eventually oscillate erratically between increasingly large *negative and positive* swings. Let's extend the series out to *N* = 1000 values to see that:

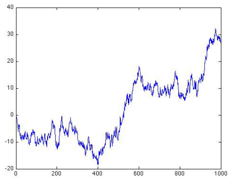


Figure 2.12(4): *N*=1000 years 'random walk'.

The swings are getting wider and wider, and they are occurring in both the positive and negative direction. Eventually, the amplitude of the swings will become arbitrarily large, i.e., infinite, even though the series will remain centered about a mean value of zero. This is an example of what we refer to in statistics as a *pathological* case.

Now let's look at what the original *N* = 200 long pure white noise series look like when there is a simple linear trend of 0.5 degree/century added on:

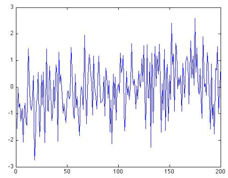


Figure 2.12(5). *N*=200 years of Gaussian White Noise with linear trend added.

Can you see a trend? In what direction? Is there a simple way to determine whether there is indeed a trend in the data that is distinguishable from random noise. That is our next topic.

Review of Basic Statistical Analysis Methods for Analyzing Data - Part 2

Establishing Trends

Various statistical hypothesis tests have been developed for exploring whether there is something more interesting in one or more data sets than would be expected from the chance fluctuations Gaussian noise. The simplest of these tests is known as *linear regression* or *ordinary least squares*. We will not go into very much detail about the underlying statistical foundations of the approach, but if you are looking for a decent tutorial [34], you can find it on Wikipedia.

The basic idea is that we test for an alternative hypothesis that posits a linear relationship between the *independent variable* (e.g., time, *t* in the past examples, but for purposes that will later become clear, we will call it *x*) and the *dependent variable* (i.e., the hypothetical temperature anomalies we have been looking at, but we will use the generic variable *y*).

The underlying statistical model for the data is:

$$y_i = a + b \cdot \chi_i + \varepsilon_i$$

(5)

where i ranges from 1 to *N, a* is the intercept of the linear relationship between *y* and *x*, *b* is the slope of that relationship, and ε is a random noise sequence. The simplest assumption is that ε is Gaussian white noise, but we will be forced to relax that assumption at times.

Linear regression determines the best fit values of *a* and *b* to the given data by minimizing the *sum of the squared differences* between the observations *y* and the values predicted by the linear model $\hat{y} = a + bx$. The *residuals* are our estimate of the variation in the data that is not accounted for by the linear relationship, and are defined by

$$\varepsilon_i = y_i - \hat{y}_i$$

(6)

For simple linear regression, i.e., *ordinary least squares*, the estimates of *a* and *b* are readily obtained:

$$b = \frac{[N \cdot \Sigma y_i x_i - \Sigma y_i \cdot \Sigma x_i]}{\left[N \cdot \Sigma {x_i}^2 - \Sigma (x_i)^2\right]}$$

(7)

and

$$a = \left({}^{1}\!/_{N}\right) \cdot \Sigma y_i - \frac{b}{N \cdot \Sigma x_i}$$

(8)

The parameter we are most interested in is *b*, since this is what determines whether or not there is a significant linear relationship between *y* and *x*.

The sampling uncertainty in *b* can also be readily obtained:

$$\sigma_b = \frac{std\,(\varepsilon)}{\left[\Sigma (x_i - \mu\,(x))^2\right]^{\frac{1}{2}}}$$

(9)

where std(ε) is standard deviation of ε and μ is the mean of *x*. A statistically significant trend amounts to the finding that *b* is significantly different from zero. The 95% confidence range for *b* is given by $b \pm 2\,\sigma_b$. If this interval does not cross zero, then one can conclude that *b* is significantly different from zero. We can alternatively measure the significance in terms of the *linear correlation coefficient*, *r* , between the independent and dependent variables which is related to *b* through

$$r = b \cdot \frac{std\,(x)}{std\,(y)}$$

(10)

r is readily calculated directly from the data:

$$r = \frac{\left({}^{1}\!/_{N-1}\right) \cdot \Sigma\,(x - \bar{x})\,(y - \bar{y})}{std\,(x) \cdot std\,(y)}$$

(11)

where over-bar indicated the mean. Unlike *b*, which has dimensions (e.g., °C per year in the case where *y* is temperature and *x* is time), *r* is conveniently a dimensionless number whose absolute value is between 0 and 1. The larger the value of *r* (either positive or negative), the more significant is the trend. In fact, the square of *r* (r^2) is a measure of the fraction of variation in the data that is accounted for by the trend.

We measure the significance of any detected trends in terms of a a *p*-value. The *p*-value is an estimate of the probability that we would wrongly reject the *null hypothesis* that there is no trend in the data in favor of the *alternative hypothesis* that there is a linear trend in the data — the *signal* that we are searching for in this case. Therefore, the smaller the *p* value, the less likely that you would observe as large a trend as is found in the data from random fluctuations alone. By convention, one often requires that $p<0.05$ to conclude that there is a significant trend (i.e., that only 5% of the time should such a trend have occurred from chance alone), but that is not a magic number.

The choice of *p* in statistical hypothesis testing represents a balance between the acceptable level of *false positives* vs. *false negatives*. In terms of our example, a false positive would be detecting a statistically significant trend, when, in fact, there is no trend; a false negative would be concluding that there is no statistically significant trend, when, in fact, there is a trend. A lower threshold (that is, higher *p*-value, e.g., $p = 0.10$) makes it more likely to detect a real but weak signal, but also more likely to falsely conclude that there is a real trend when there is not. Conversely, a higher threshold (that is, lower p-value, e.g., $p = 0.01$) makes false positives less likely, but also makes it less likely to detect a weak but real signal.

There are a few other important considerations. There are often two different alternative hypotheses that might be invoked. In this case, if there is a trend in the data, who is to say whether it should be positive ($b > 0$) or negative ($b < 0$)? In some cases, we might want only to know whether or not there is a trend, and we do not care what sign it has. We would then be invoking a *two-sided hypothesis*: is the slope *b* large enough in magnitude to conclude that it is significantly different from zero (whether positive or negative)? We would obtain a *p*-value based on the assumption of a two-sided hypothesis test. On the other hand, suppose we were testing the hypothesis that temperatures were warming due to increased greenhouse gas concentrations. In that case, we would reject a negative trend as being unphysical — inconsistent with our *a priori* understanding that increased greenhouse gas concentrations should lead to significant *warming*. In this case, we would be invoking a one-sided hypothesis. The results of a one-sided test will double the significance compared with the corresponding two-sided test, because we are throwing out as unphysical half of the random events (chance negative trends). So, if we obtain, for a given value of *b* (or *r*) a *p*-value of $p = 0.1$ for the two-sided test, than the *p*-value would be $p = 0.05$ for the corresponding one-sided test.

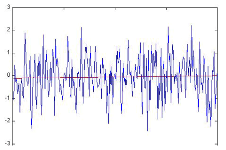
There is a nice online calculator [35], courtesy of Vassar college, for obtaining a *p*-value (both one-sided and two-sided) given the linear correlation coefficient, *r* , and the length of the data series, *N*. There is still one catch, however. If the residual *series* ε of equation 6 contains *autocorrelation*, then we have to correct the degrees of freedom, *N'*, which is less than the nominal number of data points, *N*. The correction can be made, at least approximately in many instances, using the *lag-one autocorrelation coefficient*. This is simply the linear correlation coefficient, r_1, between ε, and a carbon copy of ε lagged by one time step. In fact, r_1 provides an approximation to the parameter **ρ** introduced in equation 2. If r_1 is found to be positive and statistically significant (this can be checked using the online link provided above), then we can conclude that there is a statistically significant level of *autocorrelation* in our residuals, which must be corrected for. For a series of length $N = 100$, using a one-sided significant criterion of $p = 0.05$, we would need $r_1 > 0.17$ to conclude that there is significant autocorrelation in our residuals.

Fortunately, the fix is very simple. If we find a positive and statistically significant value of r_1, then we can use the same significance criterion for our trend analysis described earlier, except we have to evaluate the significance of the value of *r* for our linear regression analysis (not to be confused with the autocorrelation of residuals r_1) using a reduced, effective degrees of freedom *N'*, rather than the nominal sample size *N*. Moreover, *N'* is none other than the *N'* given earlier in equation 3 where we equate

$$\rho = r_1$$

That's about it for **ordinary least squares** (**OLS**), the main statistical tool we will use in this course. Later, we will encounter the more complicated case where there may be *multiple independent variables*. For the time being, however, let us consider the problem of trend analysis, returning to the synthetic data series discussed earlier. We will continue to imagine that the dependent variable (*y*) is temperature *T* in °C and the independent variable (*x*) is time *t* in years.

First, let us calculate the trend in the original Gaussian white noise series of length *N* = 200 shown in Figure 2.12(1). The linear trend is shown below:



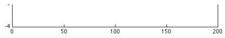


Figure 2.12(6): *N*=200 years of Gaussian White Noise with linear trend shown.

The trend line is given by: $\widehat{T} = 0.0006 \cdot t - 0.1140$, and the regression gives r = 0.0332. So there is an apparent positive warming trend of 0.0006 °C per year, or alternatively, 0.06 °C per century. Is that statistically significant? It does not sound very impressive, does it? And that r looks pretty small! But let us be rigorous about this. We have N = 200, and if we use the online calculator link provided above, we get a p-value of 0.64 for the (default) two-sided hypothesis. That is huge, implying that we would be foolish in this case to reject the null hypothesis of no trend. But, you might say, we were looking for *warming*, so we should use a one-sided hypothesis. That halves the p-value to 0.32. But that is still a far cry from even the least stringent (e.g., p = 0.10) thresholds for significance. It is clear that there is no reason to reject the null hypothesis that this is a random time series with no real trend.

Next, let us consider the red noise series of length N = 200 shown earlier in Figure 2.12(2).

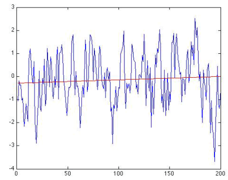


Figure 2.12(7). *N*=200 years of Gaussian 'red noise' with ρ=0.6 with linear trend shown.

As it happens, the trend this time appears nominally greater. The trend line is now given by: $\widehat{T} = 0.0014 \cdot t - 0.2875$, and the regression gives r = 0.0742. So, there is an apparent positive warming trend of 0.14 degrees C per century. That might not seem entirely negligible. And for N = 200 and using a one-sided hypothesis test, r = 0.0742 is statistically significant at the p = 0.148 level according to the online calculator. That does not breach the typical threshold for significance, but it does suggest a pretty high likelihood (15% chance) that we would err by not rejecting the null hypothesis. At this point, you might be puzzled. After all, we did not put any trend into this series! It is simply a random realization of a red noise process.

Self Check

So why might the regression analysis be leading us astray this time?

Click for answer.

The problem is that our residuals are not uncorrelated. They are red noise. In fact, the residuals looks a lot like the original series itself:

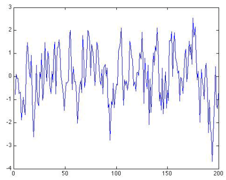


Figure 2.12(8). Residuals from linear regression with *N*=200 years of Gaussian 'red noise' with ρ=0.6

This is hardly coincidental; after all, the trend only accounts for $r^2 = 0.0742^2 = 0.0055$, i.e., only about half a percent, of the variation in the data. So 99.5% of the variation in the data is still left behind in the residuals. If we calculate the lag-one autocorrelation for the residual series, we get r_1 = 0.54. That is, again not coincidentally, very close to the value of ρ = 0.6 we *know* that we used in generating this series in the first place.

How do we determine if this autocorrelation coefficient is statistically significant? Well, we can treat it like it were a correlation coefficient. The only catch is that we have to use *N-1* in place of *N*, because there are only *N-1* values in the series when we offset it by one time step to form the lagged series required to estimate a lag-one autocorrelation.

Self Check

Should we use a one-sided or two-sided hypothesis test?

Click for answer.

If we use the online link and calculate the statistical significance of r_1 = 0.54 with *N-1* = 199, we find that it is statistically significant at p < 0.001. So, clearly, we cannot ignore it. We have to take it into account.

So, in fact, we have to treat the correlation from the regression r = 0.074 as if it has $N' = (1 - 0.54) / (1 + 0.54)\, 200 = 0.30\,(200) = 59.9 \approx$ 60 degrees of freedom, rather than the nominal N = 200 degrees of freedom. Using the interactive online calculator, and replacing N = 200 with the value N' = 60, we now find that a correlation of r = 0.074 is only significant at the p = 0.57 (p = 0.29) for a two-sided (one-sided) test, hardly a level of significance that would cause us to seriously call into doubt the null hypothesis.

At this point, you might be getting a bit exasperated. When, if ever, *can* we conclude there is a trend? Well, why don't we now consider the case where we know we added a real trend in with the noise, i.e., the example of Figure 2.12(5) where we added a trend of 0.5°C/century to the Gaussian white noise. If we apply our linear regression machinery to this example, we do detect a notable trend:

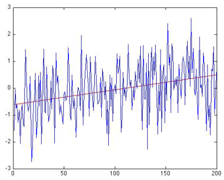


Figure 2.12(9). N=200 years of Gaussian white noise with added linear trend of 0.5 degrees/century; the red line shows trend recovered by the linear regression.

Now, *that's* a trend - your eye isn't fooling you. The trend line is given by: $\widehat{T} = 0.0056 \cdot t - 0.619$. So there is an apparent positive warming trend of 0.56 °C per century (the 95% uncertainty range that we get for b, i.e., the range $b \pm 2\,\sigma_b$, gives a slope anywhere between 0.32 and 0.79 °C per century, which of course includes the true trend (0.5 °C/century) that we know we originally put in to the series!). The regression gives r = 0.320. For N = 200 and using a one-sided hypothesis test, r = 0.320 is statistically significant at p<0.001 level. And if we calculate the autocorrelation in the residuals, we actually get a small *negative* value ($r_1 = -0.095$), so autocorrelation of the residuals is not an issue.

Finally, let's look at what happens when the same trend (0.5 °C/century) is added to the random red noise series of Figure 2.12(2), rather than the white noise series of Figure 2.12(1). What result does the regression analysis give now?

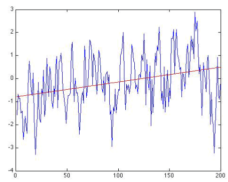


Figure 2.12(10). N=200 years of Gaussian red noise with ρ=0.6 and added linear trend of 0.5 degrees/century; the red line shows trend recovered by the linear regression.

We still recover a similar trend, although it's a bit too large. We know that the true trend is 0.5 degrees/century, but the regression gives: $\widehat{T} = 0.0064 \cdot t - 0.793$. So, there is an apparent positive warming trend of 0.64 °C per century. The nominal 95% uncertainty range that we get for b is 0.37 to 0.92 °C per century, which again includes the true trend (0.5 degrees C/century). The regression gives r = 0.315. For N = 200 and using a one-sided hypothesis test, r = 0.315 is statistically significant at the p < 0.001. So, are we done?

Not quite. This time, it is obvious that the residuals will have autocorrelation, and indeed we have that r_1 = 0.539, statistically significant at p < 0.001. So, we will have to use the reduced degrees of freedom N'. We have already calculated N' earlier for ρ = 0.54, and it is roughly N' = 60. Using the online calculator, we now find that the one-sided p = 0.007, i.e., roughly p = 0.01, which corresponds to a 99% significance level. So, the trend is still found to be statistically significant, but the significance is no longer at the astronomical level it was was when the residuals were uncorrelated white noise. The effect of the "redness" of the noise has been to make the trend less statistically significant, because it is much easier for red noise to have produced a spurious apparent trend from random chance alone. The 95% confidence interval for b also needs to be adjusted to take into account the autocorrelation, though just how to do that is beyond the scope of this course.

Often, residuals have so *much* additional structure — what is sometimes referred to as *heteroscedasticity* (how's that for a mouthful?) — that the assumption of simple autocorrelation is itself not adequate. In this case, the basic assumptions of linear regression are called into question and any results regarding trend estimates, statistical significance, etc., are suspect. In this case, more sophisticated methods that are beyond the scope of this course are required.

Now, let us look at some *real temperature data*! We will use our very own custom online Linear Regression Tool [36] written for this course. The demonstration how to use this tool has been recorded in three parts below (click each link to open a new window and then the arrow to begin the demonstration):

Part 1 [37]

Part 2 [38]

Part 3 [39]

You can play around with the temperature data set used in this example using the Linear Regression Tool [36]

Review of Basic Statistical Analysis Methods for Analyzing Data - Part 3

Establishing Relationships Between Two Variables

Another important application of OLS is the comparison of two different data sets. In this case, we can think of one of the time series as constituting the independent variable x and the other constituting the independent variable y. The methods that we discussed in the previous section for estimating trends in a time series generalize readily, except our predictor is no longer time, but rather, some variable. Note that the correction for autocorrelation is actually somewhat more complicated in this case, and the details are beyond the scope of this course. As a general rule, *even* if the residuals show substantial autocorrelation, the required correction to the statistical degrees of freedom (N'), will be small as long as either one of the two time series being compared has low autocorrelation. Nonetheless, any substantial structure in the residuals remains a cause for concern regarding the reliability of the regression results.

We will investigate this sort of application of OLS with an example, where our independent variable is a measure of El Niño — the so-called *Niño 3.4 index* — and our dependent variable is December average temperatures in State College, PA.

The demonstration is given in three parts below (click each link to open a new window and then the arrow to begin the demonstration):

Demonstration - Part 1 [40]

Demonstration - Part 2 [41]

Demonstration - Part 3 [42]

You can play around with the data set used in this example using this link: Explore Using the File testdata.txt [43]

Problem Set #1

Activity: Statistical Analysis of Climate Data

NOTE: For this assignment, you will need to record your work on a word processing document. Your work must be submitted in Word (.doc or .docx), or PDF (.pdf) format.

For this activity, you will use the application below to perform basic statistical analyses of climate data. The data we will use are global temperature anomalies and Niño 3.4 index, both measured in °C. You need to:

- determine historical trends in global temperatures and determine if there has been an increase in the trend.
- analyze the influence of El Niño on global temperatures.

Link to Linear Regression Tool [44]

Directions

1. First, save the Problem Set #1 Worksheet [45] to your computer. You will use this word processing document to electronically record your work in the remaining steps.
    - Save the worksheet to your computer by right-clicking on the link above and selecting "Save link as..."
    - The worksheet is in Microsoft Word format. You can use either Word or Google Docs (free) to work on this assignment. You will submit your worksheet at the end of the activity, so it must be in Word (.doc or .docx) or PDF (.pdf) format so the instructor can open it.
    - Please show your work! When you are explicitly asked to create plots in a question, please cut-and-paste graphics and the output from the screen (e.g., by first printing the output to a pdf file and then directly inserting into the worksheet) to submit along with your discussion and conclusions.
2. Use the plotting tool to create a line plot of global temperature anomalies vs. time over the full 160 year period (1850 - 2009). Determine the basic statistics of the time series, i.e., the mean and standard deviation. If the mean is not zero, can you guess the 30 year base period that was used to calculate the anomalies? To do that, note that by definition, the mean of the base period used to calculate anomalies should be zero. Visually determine the year where the temperature anomalies graph crosses zero. Zoom in on different 30-year intervals around that year and use Viewport Data under the Statistics tab to view the mean for each selected 30-year sub-set of data, until you find a period for which the mean is closest to zero.
3. Evaluate the linear trend in global temperature anomalies over the full 160 year period (1850 - 2009) following the steps below. (A) Add a trend line to the plot of global temperature anomalies vs. time over the full 160 year period: use the Trend Lines tab. Determine the slope of the linear regression line, b, in ºC/century and the correlation coefficient, r. As a side exercise, calculate the overall warming trend in temperature over the 1850-2009 period by multiplying the slope by the number of years in the period. (B) Assess statistical significance of the linear trend in global temperature anomalies without considering autocorrelation: use the online Statistical Calculator tool [35] from Lesson 2, Statistical Analysis Part 2 to calculate the p-value for the number of samples, N, and the correlation coefficient, r, from part A. Interpret the p-value. Use the standard error of the slope, Sb, to calculate 95% confidence interval for the slope as b ±2Sb, and report the calculated warming range in ºC/century. (C) Determine lag-one autocorrelation coefficient for the residuals, ρ. To do that, run the regression model using the Regression Model tab: select Model Parameters = year, Target Observation = Temp Anom. Plot the residuals: in the Plot Settings tab, select year for X and Model Residuals for Y. Check whether the autocorrelation is statistically significant using the Statistical Calculator. Remember that lag-one autocorrelation is simply correlation between the original data and an exact copy of the data but shifted by one time step, so the number of samples, N, is decreased by one; and the correlation coefficient, r, should be substituted by autocorrelation coefficient, ρ. Interpret the p-value. (D) Reassess statistical significance of the linear trend in the global temperature anomalies, taking into account autocorrelation of the residuals. Remember that in the presence of a significant autocorrelation, the actual number of samples, N, must be replaced by the degrees of freedom, N': use formula (2) from Lesson 2 to calculate N'. Use the Statistical Calculator to calculate the p-value for N = N' and r from part (A). Interpret the p-value. (E) Look closely at the plot of the residuals that you created in part (C). Do you see evidence of heteroscedasticity (additional structure superimposed on the random fluctuations)? Do you think that the hypothesis of a simple linear warming trend in this data series is appropriate or not?
4. Test the hypothesis that there is a difference in trend over two sub-periods: the first 130 years and the final 30 years. Follow the steps below. (A) Use the Plot Settings tab to plot the entire data series and zoom in to select 1850-1979 sub-period. Select Viewport in the Trend Lines tab to perform linear regression for the sub-period. Determine the slope of the linear regression line, b, in ºC/century and the correlation coefficient, r. Use the standard error of the slope, Sb, to calculate 95% confidence interval for the slope in ºC/century. (B) Repeat the analysis for the 1980-2009 sub-period. (C) Determine whether trends are statistically different: you need to check whether the 95% confidence intervals for the two sub-periods overlap. Based on your results, has global warming accelerated over the past 30 years? What important caveat about the 95% confidence intervals was not taken into account in our analysis?
5. Is there a statistically significant influence of El Niño on global temperatures? (A) Use the Plot Settings tab to plot global temperature anomalies (Temp Anom) over time (Plot #1) and then Niño 3.4 index (Niño) over time (Plot #2) on the same plot. Determine the number of years, N, in the time interval over which the two data series overlap. This is the number of samples you will use in the analyses below. (B) Plot the relationship between Niño 3.4 index and global temperature anomaly: use the Plot Settings tab to plot Niño 3.4 index (Niño) on the X-axis and global temperature anomalies (Temp Anom) on the Y-axis. Determine the slope of the linear regression line, b (in °C change in global temperature per a unit change in Niño 3.4 index) and the correlation coefficient, r. Assess statistical significance of the linear trend without accounting for autocorrelation. (C) Plot model residuals over time and discuss whether heteroscedasticity of the regression model residuals is a point of concern in your analysis.
6. Given that there are La Niña year conditions this year, how do you expect ENSO to influence this year's global temperature? ENSO – El Niño-Southern Oscillation – is a climate pattern of oscillation between El Niño events (the positive phase, i.e., Niño 3.4 index above average) and La Niña events (the negative phase, i.e., Niño 3.4 index below average). Using the most negative observed value of the Niño 3.4 index during the historical time period used for the regression equation from question (5), determine the temperature anomaly predicted by the regression model for this most negative value. By comparing this temperature anomaly to the temperature anomaly expected in neutral (i.e., Niño 3.4 index = 0) conditions, estimate the largest perturbation of global temperatures expected in the most extreme La Niña events.
7. Save your word processing document as either a Microsoft Word or PDF file in the following format: PS1_AccessAccountID_LastName.doc (or .pdf).

    For example, student Elvis Aaron Presley's file would be named "PS1_eap1_presley.doc". This naming convention is important, as it will help the instructor match each submission with the right student!

Submitting your work

Upload your file to the "Problem Set #1" assignment in Canvas by the due date indicated in the Syllabus.

Grading rubric

The instructor will use the general grading rubric for problem sets [46] to grade this activity.

Solutions will be uploaded to *Lesson 2: Climate Observations, Part 1* in *Canvas* after the due date.

Lesson 2 Summary

In this lesson, we reviewed key observations that detail how our atmosphere and climate are changing. We have seen that:

- greenhouse gas concentrations, including atmospheric CO_2 and methane, are increasing dramatically and these increases are associated with human activity;
- the surface of the Earth is warming and certain regions (e.g., the Arctic) are warming faster than others, consistent, as we will see, with expectations from climate model projections;
- the vertical pattern of the warming indicates that the surface and lower atmosphere (troposphere) are warming, while the atmosphere is cooling at altitude (in the stratosphere), a pattern that is consistent with greenhouse warming, but not with the natural factors such as solar output changes;
- there is a complicated pattern of changes in rainfall patterns around the globe, with some regions becoming wetter while other regions become drier;
- despite the heterogeneous pattern of changes in rainfall, there is a trend towards more widespread drought, consistent with the additional impact of warming on evaporation from soil.

We also learned how to analyze basic relationships in observational data, including:

- how to assess whether or not there is a statistically significant trend over time in a data series;
- how to assess whether or not there is a statistically significant relationship between two distinct data series.

In our next lesson, we will look at some additional types and sources of observational climate data, and we will explore some additional tools for analyzing data.

Reminder - Complete all of the lesson tasks!

You have finished Lesson 2. Double-check the list of requirements on the first page of this lesson to make sure you have completed all of the activities listed there before beginning the next lesson.

Lesson 3 - Climate Observations, part 2

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

About Lesson 3

In Lesson 2, we focused on atmospheric observations documenting historical changes in the climate system. In this lesson, we will turn to other evidence of climate change, including paleoclimate data that can be used to document, albeit with added uncertainty, more distant past changes in climate, and other variables documenting changes in the climate system including measures of ocean circulation, changes in sea ice, glaciers, and other climate change data documenting extreme weather, including tropical cyclone and hurricanes.

What will we learn in Lesson 3?

By the end of Lesson 3, you should be able to:

- discuss the various modern observational and paleoclimate data sets relevant to assessing modern-day climate change, and their uncertainties;
- discuss the role of both the oceans and atmosphere in observed climate variability and climate change;
- perform statistical analyses where there are multiple potential factors influencing some climate variable.

What will be due for Lesson 3?

Please refer to the Syllabus for specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 3. Detailed directions and submission instructions are located within this lesson.

- Read:
    - IPCC Fifth Assessment Report, Working Group 1 [2]
    - Summary for Policy Makers [3]
        - B.2 Ocean: p. 8
        - B.3 Cryosphere: p. 9-10
        - B.4 Sea Level: p. 11
    - Dire Predictions, v.2: p. 36-37, 100-101, 110-111, 148-149
- Problem Set #2: Analyze the statistical relationships between different natural and human factors and global temperature variations over the past 1000 years.
- Take Quiz #1.

Questions?

If you have any questions, please post them to our *Questions?* discussion forum (not e-mail), located under the Home tab in Canvas. The instructor will check that discussion forum daily to respond. Also, please feel free to post your own responses if you can help with any of the posted questions.

Sea Ice, Glaciers, Ice Sheets, and Global Sea level

From the standpoint of climate change impacts, nothing could be more important than the potential changes in Earth's cryosphere — that is, the sea ice, the glaciers, and the two major ice sheets.

Sea Ice

As temperatures warm in the Arctic, the extent of summer sea ice coverage continues to decrease. Ice extent dropped to precipitous levels in 2007. Arctic sea ice seemed to recover in 2008, but then the sea-ice cover decline resumed. In 2012, the area of Arctic sea ice at the end of the summer melting season reached a low of 3.3 million square kilometers (1.3 million square miles), well below the projections of IPCC models.

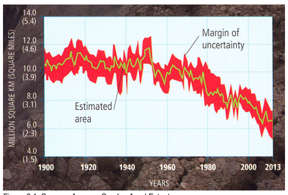


Figure 3.1: Summer Average Sea Ice Areal Extent.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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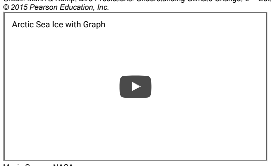
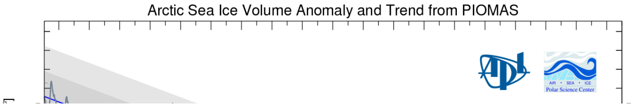


Movie Source: NASA [47]

Perhaps more significantly, much of the more resilient, thicker multi-year ice (the ice that survives the summer melt season so that it can further accumulate winter after winter) has disappeared, and the remaining ice is largely just seasonal ice that is far more prone to melting. In fact, when the decreasing thickness as well as extent is taken into account, based on sophisticated computer analyses, the decrease in sea ice volume (the most relevant quantity) is in fact seen to be declining even more abruptly than the extent alone.



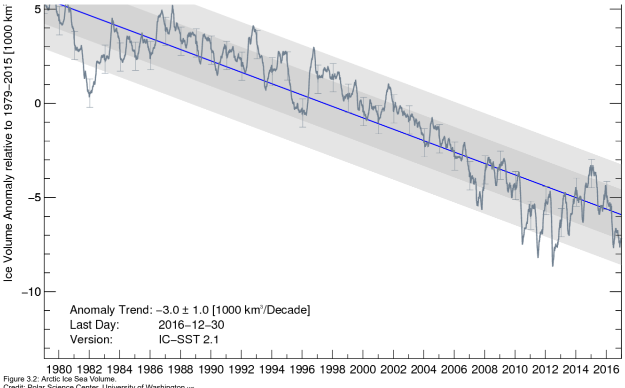


Figure 3.2: Arctic Ice Sea Volume.
Credit: Polar Science Center, University of Washington [48]

As we will see later, the decreases shown by the observations are actually ahead of schedule as far as state-of-the-art climate change projections are concerned. In fact, some predictions based on the observed trends have Arctic sea ice disappearing completely during the summer in as few as a couple decades.

Glaciers

Mountain glaciers can be found on all of the continents of the world with the exception of Australia. They exist typically at high elevations where the accumulation of snow out-paces the *ablation* — the loss of ice through *melting* or *sublimation*. Because glaciers are vertically distributed, the accumulation and ablation may take place in different locations. For example, the accumulation may take place largely at the apex of a mountain where conditions are cold and most if not all precipitation falls as snow, while the ablation may take place at the periphery of the glacier at lower elevation, where temperatures are high enough, at least seasonally, for ice to melt.

Mountain glaciers have been retreating around the world over the past century. Below are some examples of "before and after" photos demonstrating the dramatic retreat of mountain glaciers in various regions of the world including (top) the McCall Glacier of the Brooks Range in Alaska, (middle) Muir Glacier in Alaska, and (bottom) Qori Kalis Glacier in Peru.

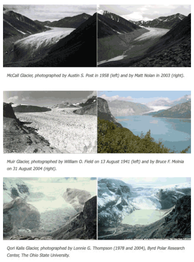


Figure 3.3: Documented Retreat of Various Mountain Glaciers Over the Past Century.
Credit: National Snow and Ice Data Center [49]

This is primarily because of warming temperatures leading to increased summer melt.

In some cases, the situation is a bit more complicated. Consider for example Mount Kilimanjaro. The iconic glacier fields atop this mountain are essentially located at the equator. This iconic equatorial ice cap was immortalized by Ernest Hemingway during the 1930s in his novel The Snows of Kilimanjaro [50]. Ironically, the snows of Kilimanjaro are disappearing rapidly (see below).

Figure 3.4: Retreat of Kilimanjaro's Ice Fields over the Past Century.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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At the current rate of ablation, Kilimanjaro's ice fields will be gone within the next two decades. Is this imminent disappearance due to global warming? Well, Kilimanjaro's ice fields have been around for at least 12,000 years. It would seem quite a coincidence, therefore, that they just happened to choose to disappear now. However, it isn't quite as simple as one might imagine. Melting is probably not the primary process responsible for the ablation of the ice. At this altitude of nearly 6 km (placing it in the mid-troposphere), the primary mechanism by which ice is lost is sublimation, not melting. Moreover, as with many tropical glaciers (see e.g., this article on Tropical Glacier Retreat [51] at the site RealClimate.org), changes over time in the overall mass balance are heavily influenced by accumulation (i.e., the amount of snowfall), not just melting. Some have argued, for these reasons, that the rapid disappearance of Kilimanjaro's ice cap cannot be blamed on human-caused climate change. That argument is probably wrong, however. While changes in humidity and precipitation have clearly played a role in decreasing accumulation, these changes may in part reflect the large-scale reorganization of the atmospheric circulation that is tied to human-caused climate change. Moreover, some direct melting of Kilimanjaro's ice fields has been observed in recent decades, and that melting is almost certainly part of the picture. Such increases in melting have been observed for high-elevation mountain glaciers throughout the tropics, and are tied to a large-scale warming of the tropical mid-troposphere that appears to be connected with the larger-scale pattern of a warming troposphere.

Now, there *are* some exceptions to the pattern of widespread global glacial retreat. See the map below. Certain outlet glaciers in Scandinavia and the Pacific Northwest have actually *expanded* in extent in recent decades.

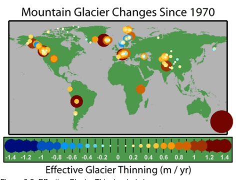

Figure 3.5: Effective Glacier Thinning (m/yr).
Credit: Image created by Robert A. Rohde / Global Warming Art [52]

Think About It!

Why do you suppose that expansion is taking place in some locations? *[Hint: what do those regions have in common?*]

Click for answer.

If we aggregate all of the major glaciers around the world into a single estimate of 'mass balance', i.e., the net change in ice mass which represents the balance between accumulation through snowfall and loss through melting and sublimation, we find a pronounced trend towards decreasing glacial ice mass, which in many respects mirrors the loss of sea ice shown earlier. However, unlike melting sea ice which does not contribute to global sea level rise (because the ice is already floating on the ocean), the melting glaciers *do* make a significant contribution to rising global sea levels — a topic we address below.

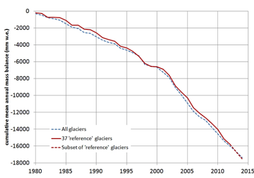

Figure 3.6: Mean cumulative mass balance of all glaciers in recent decades.
Credit: World Glacier Monitoring Service [53]

Ice Sheets

While we have seen that the world's glaciers are melting *en masse*, and at an accelerating rate, we have not yet addressed the behavior of the two largest glaciers in the world — glaciers that are so large, we call them *continental ice sheets*. There are two continental ice sheets — the Greenland ice sheet and the Antarctic Ice Sheet. In reality, only a portion of the Antarctic ice sheet is susceptible to collapse. The East Antarctic ice sheet sits at a relatively high elevation and is relatively stable. It is unlikely to disappear under most projected climate change scenarios. However, the lower elevation *West Antarctic* ice sheet is likely susceptible to mass wastage.

The mass of the major ice sheets, like that of smaller glaciers, depends on the balance between accumulation and loss to ablation. Also, as with smaller glaciers, the regions of accumulation and ablation are typically not the same. The primary accumulation is in the colder interiors of the continents, while ice flowing out towards the periphery at lower latitudes is subject to melting and the calving of ice into the ocean. In the case of the Greenland ice sheet, fissures known as *moulins* may form, allowing meltwater to percolate to the bottom and help lubricate streams of melting ice that escape to the ocean in ice channels. In the case of the Antarctic Ice sheet, ice calves into the southern ocean at the periphery of expansive ice shelves that extend out over the relatively warm ocean.

Because of the complex balance between the processes favoring accumulation and ablation, it was not known for some time whether the observed warming of the globe had in fact led to any net loss of ice mass for either of the two continental ice sheets. In recent years, however, careful satellite measurements have suggested that detectable changes are indeed under way. The area of summer ablation over Greenland, for example, has expanded greatly in recent years (see figure below).

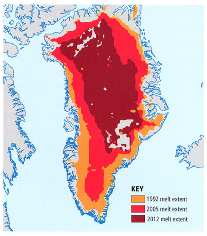

Figure 3.7: Greenland's Melting Continental Ice Sheet.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Detailed satellite measurements using a variety of techniques based on *altimetry* and the measurement of *gravitational anomalies* suggest that ablation is now outpacing accumulation for both the Greenland and Antarctic ice sheets. In other words, like the vast majority of small glaciers worldwide, the two continental ice sheets themselves are now losing mass — and, as discussed below, contributing to sea level rise.

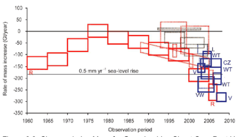

Figure 3.8: Changes in Ice Mass for Greenland Ice Sheet Over Past Half Century.
Credit: The Copenhagen Diagnosis Report [54]

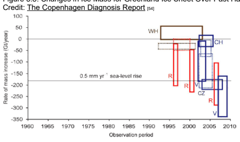

Figure 3.9: Changes in Ice Mass for the Antarctic Ice Sheet over Past Half Century.
Credit: The Copenhagen Diagnosis Report [54]

Global Sea Level

We have seen that the world's ocean surface is warming [55]. Indeed, as we will see later in this lesson, that warmth is slowly penetrating down into the deep ocean. As ocean water warms, it expands, and thus contributes to raising sea level. We refer to this component of global sea level rise as the *thermosteric* component. But there are other key contributions of global warming to global sea level. In fact, we've just discussed them above: the melting of glaciers, and the loss of ice mass now underway for both of the two continental ice sheets.

The latter contribution exceeds the expectations scientists had just a few years ago, before there was any consensus that the decay of the Greenland and West Antarctic ice sheets was likely to happen in the near future, let alone already be underway. Not surprisingly, we are observing that global sea level rise is proceeding at the very upper extreme of the range that was projected by the models just a couple decades ago (see below).

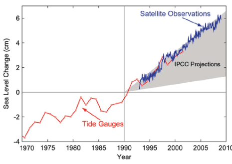

Figure 3.10: Observed Sea Level Rise over Past Half Century Compared with 1990 IPCC Projections.
Credit: The Copenhagen Diagnosis Report [54]

The Oceans

As we have seen previously, the entire surface of the Earth has warmed by a bit less than 1°C over the past century. It is evident that the oceans, on average, have warmed a bit less than the land regions. The warming of the oceans has been damped by their greater *thermal inertia* — water has a greater heat capacity than land, and the oceans are several km deep, and heat can efficiently be buried below the surface.

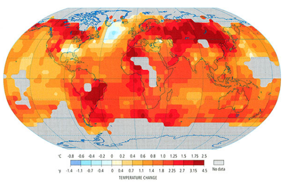

Figure 3.11: Trends in Global Surface Temperature 1901 - 2012.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Part of the reason that the surface of the oceans is warming less than the land surface, then, is that a good deal of the heat is being buried below the ocean surface. In other words, the heating from global warming is slowly diffusing down through the ocean, warming the entire layer of ocean water several km down below the surface.

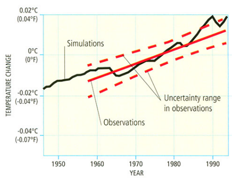

Figure 3.12: Actual vs. simulated deep-ocean temperature changes.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Because the processes (small-scale convection and diffusion) by which this heat penetrates downward through the ocean are slow, these processes lead to a substantial delay in how the ocean (surface and sub-surface) warms in response to any given radiative forcing, including that due to increasing greenhouse gas concentrations. These considerations have profound implications for global sea level rise, as alluded to in the previous section on sea level rise. Because sea water expands with warming, ocean levels continue to slowly rise as the heating penetrates down through the deep ocean. As we will see later in the course when we discuss climate change impacts [56], it is such slow, delayed responses to global warming that leads to very long-term consequences of policy decisions being made today; global sea level will continue to rise for several centuries, even if we were to freeze greenhouse gas concentrations at current levels. Such lasting impacts of current human influences on climate are referred to as the *committed climate change*. They are part of the motivation for calls by many for immediate reductions in greenhouse gas emissions.

It is important to recognize that the oceans are not simply a passive reservoir for absorbing surface heating. They are, as we saw in our first lesson, highly dynamic [57]. Oceans play a key role, for example, in transporting heat from low latitudes to higher latitudes to help relieve the imbalance in solar heating. Much of this transport takes place within the horizontal ocean gyres. As the gyres are governed primarily by the latitudinal pattern of variations in surface winds, they are relatively robust. Climate change may alter prevailing wind patterns somewhat, but the main features, e.g., the presence of easterly surface winds in the tropics and westerly surface winds in mid-latitudes, are not projected to change. Therefore, we don't expect substantial changes in the role played in the climate system by the horizontal ocean gyres.

There could be a larger role, however, played by the ocean's thermohaline circulation.

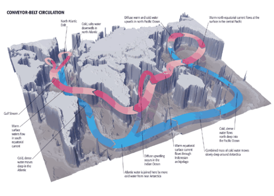

Figure 3.13: Conveyor Belt Circulation.
Credit: Mann & Kump, *Dire Predictions*

We know that the thermohaline circulation plays a significant role in the natural long-term variability of the climate system. Indeed, there is a mode of climate variability known as the *Atlantic Multidecadal Oscillation* [58] (or simply 'AMO') that appears to arise from long-term oscillations in the North Atlantic component of the thermohaline circulation. While there is some disagreement about the larger influences of this mode of natural climate variability, there is evidence that it impacts sea surface temperatures in the extratropical North Atlantic ocean and neighboring regions of North America and Europe.

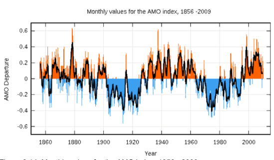

Figure 3.14: Monthly values for the AMO index, 1856 - 2009.
Credit: Wikipedia: Atlantic multidecadal oscillation [58]

There may also be a role of the thermohaline circulation in climate change itself. As we will see later in the course when we examine climate model projections of future climate change, there is the possibility that the thermohaline circulation could weaken, or even collapse, in a global warming scenario. It is a combination of the salty and cold properties of surface waters in the sub-polar North Atlantic that leads to their high density, and consequent sinking motion. That sinking motion forms the descending limb of the thermohaline circulation, so any substantial freshening and warming of these waters could inhibit that sinking. It has long been suspected that global warming, through the influx into the North Atlantic of fresh water from melting land snow and ice, could thus inhibit or even shut down the thermohaline circulation. There is quite a bit of debate as to whether or not the data show that such a weakening is underway. Direct measurements of the strength of the thermohaline circulation are scarce, and are sparse over time. Thus, evidence for trends are at best equivocal.

As we will see later in this lesson, there is some evidence that a thermohaline circulation collapse scenario may have played out at the end of the last ice age, during a period known as the *Younger Dryas*. At that time, the large amount of meltwater produced from the initial termination of the ice age appears to have shut down the thermohaline circulation. Since the thermohaline circulation is a source of poleward heat transport in regions surrounding the North Atlantic, this event appears to have temporarily sent the climate back into a glacial state before the final termination of the ice age a thousand or so years later. Could such scenario play out because of human-caused climate change? We will revisit that question in a later lecture.

The El Niño/Southern Oscillation

We have already seen an example of a natural mode of climate variability above in the AMO. However, the single most important mode of natural variability in the climate system—certainly on interannual timescales—is the El Niño/Southern Oscillation or *ENSO*. ENSO represents a coupled mode of the ocean and atmosphere, which is to say that the atmosphere influences the ocean (primarily through the impact of surface winds on horizontal and vertical 'upwelling' ocean currents), while the oceans, in turn, influence the atmosphere (primarily through the effect of east/west variations in ocean surface temperature, which, in turn, drive a pattern of circulation in the overlying atmosphere). The net result is that the tropical Pacific combined ocean/atmosphere system naturally oscillates with a characteristic timescale of roughly 2-7 years.

During the *El Niño* phase of the oscillation, the eastern/central tropical Pacific is warmer than usual. Sea level in the eastern tropical Pacific is higher than usual because the waters are warm (and thus less dense), in the absence of vertical upwelling of colder, denser water. The warmer waters in the central equatorial Pacific lead to rising motion in the atmosphere, shifting the rising limb of the so-called *Walker Circulation* (the vertical and longitudinal pattern of atmospheric circulation over the equatorial Pacific), and associated tendency for rainfall, from its normal position in the western equatorial Pacific. This circulation pattern, in turn, implies a decrease in the strength of the easterly trade winds in the eastern tropical Pacific. But it is those trade winds that are responsible for upwelling of the cold waters in the first place. Thus, the ocean and atmosphere work in concert to sustain the ENSO pattern of temperature, winds, and rainfall. The *La Niña* state, when sea surface temperatures are cooler than normal in the eastern and central tropical Pacific, is based on these various features of the tropical Pacific ocean-atmosphere system being opposite to that described above for the El Niño phase.

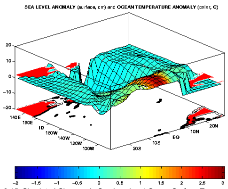

Figure 3.15: Simulated Changes in Sea Level and Ocean Surface Temperature associated with ENSO.
Credit: NOAA [59]

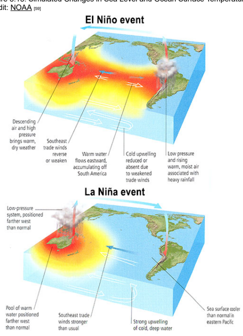

Figure 3.16: Diagrams of El Niño and La Niña Events.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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The system tends to oscillate between the El Niño and La Niña phases because of equatorial ocean wave dynamics that are beyond the scope of this course. The net effect of these waves—*Kelvin waves* and *Rossby waves*—is to make the system unstable so that it does not persist either in the El Niño or La Niña phase for more than a year or so, constantly oscillating between these two phases with a characteristic 2-7 year timescale.

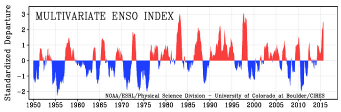

Figure 3.17: Multivariate ENSO Index.
Credit: NOAA [60]

If we look at the pattern of evolution of sea surface temperature *anomalies* over the tropical Pacific during recent years, we can clearly see this alternation between states when the eastern and central tropical Pacific are relatively warm (El Niño events) and states when the regions are relatively cold (La Niña events).

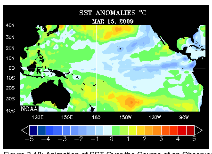

Figure 3.18: Animation of SST Over the Course of an Observed La Niña to El Niño Transition.
Credit: NOAA [61]

By some measures, ENSO is the largest signal in the observational climate record, even competing with the climate change signal itself. That is because El Niño and La Niña events can be large, leading to more than a degree C cooling or warming over a large part of the tropical Pacific ocean, and ENSO events have global *teleconnections*. That is to say, ENSO affects seasonal weather patterns around the world through the impact of the tropical Pacific cooling/warming on the extratropical jet streams, the Monsoons, and other regional atmospheric circulation patterns. As we will see in the next section, ENSO even influences Atlantic seasonal hurricane activity. The largest and most reliable regional impacts of ENSO are shown below:

Figure 3.19: Large Scale Impacts of El Niño (Northern Hemisphere Winter).
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Given the prominent impact of ENSO on year-to-year variations in climate around the world, it is well worth asking, how might ENSO itself change, in response to human-caused climate change? Such questions are key to assessing regional impacts of future climate change. Yet, we don't have the firm answers here that one might expect or want. In the time series plot above, there does seem to be some evidence of a trend towards more El Niño-like conditions in recent decades, but it is difficult to establish any statistical significance in this trend. Indeed, some climate models suggest the possibility of the opposite trend—a prevalence of the La Niña state—in response to anthropogenic climate change. We will return to the issue of climate variability [62] in a later lecture on projections of future climate change.

Tropical Cyclones / Hurricanes

There is perhaps no more impressive or—for that matter—deadly a weather phenomenon than tropical cyclones (and hurricanes—which are simply strong tropical cyclones).

We will focus, for the purpose of our discussion, on Atlantic tropical cyclones (TCs), as they are the best observed, providing records that go back more than a century (albeit with some uncertainties, particularly in earlier decades), and they are most relevant from the standpoint of North American impacts.

The 2005 Atlantic hurricane series was the most active season ever, with a total of 28 TCs forming over the course of the Atlantic hurricane season (which formally runs from Jun 1-Nov 30, though storms sometimes occur outside this window).

The 2005 hurricane season was most notable, however, for the series of extremely powerful category 5 Hurricanes (the strongest category) that made landfall along the U.S. coastline, including Katrina which devastated New Orleans in late August, Rita which also made landfall on the U.S. Gulf Coast in late September, and Wilma—the strongest Atlantic hurricane ever—which made multiple landfalls in Mexico and Florida in late October.

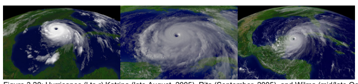
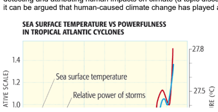


Figure 3.20: Hurricanes (l to r) Katrina (late August, 2005), Rita (September, 2005), and Wilma (mid/late October, 2005).
Credit: Wikimedia Commons

One might well ask, is the apparent recent trend towards more destructive Atlantic hurricanes in some way tied to climate change? Well—as we will discuss in more detail in the next section on *extreme weather*—it is *never* possible to attribute *any single weather event* to climate change. But we can arguably see the impacts of climate change in the longer-term shifts in frequency and intensity of such events. In the case of Atlantic tropical cyclones & hurricanes, one convenient measure of activity is the net *power dissipation*, a measure that measures the energy dissipated by tropical cyclones & hurricanes as they interact with the ocean surface, averaged over the integrated lifetime of all storms during a particular season. Simple theoretical arguments developed by MIT hurricane expert Kerry Emanuel [63] imply that this measure of hurricane activity should closely follow sea surface temperatures.

If we compare sea surface temperatures over the tropical Atlantic in the main development region (MDR) for tropical storms during the core of the hurricane season (August-October) to estimates of the *power dissipation index*, we do, indeed, find a very close relationship as far back as reliable records go (the mid 20th century). Indeed, the increase in powerfulness of storms in recent decades (the broader context for the unusually active 2005 season) does appear to be tied to the warming of Atlantic ocean surface temperatures. Moreover, computer modeling studies aimed at detecting and attributing human impacts on climate (a topic discussed in later lectures) do indeed tie this warming to human causes. In this sense, it can be argued that human-caused climate change has played a role in the increased destructiveness of Atlantic hurricanes in recent years.



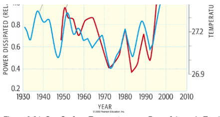


Figure 3.21: Sea Surface Temperatures vs. Powerfulness in Tropical Atlantic Cyclones.
Credit: Mann & Kump, *Dire Predictions*

What about the *number* of Atlantic TCs, including the hyperactive 2005 season with its 28 named storms? Is this part of a longer-term trend? And if so, is that trend related to human influences on the climate? This question is still being actively debated within the scientific community. For one thing, there is some question about how reliable long-term records of TC counts are, particularly prior to the use of modern aircraft reconnaissance in the mid 1940s (if you're interested in the details, you can find a journal article about this topic by Michael Mann entitled *Evidence for a modest undercount bias in early historical Atlantic tropical cyclone counts* [64]). Furthermore, while sea surface temperatures are one factor driving Atlantic tropical cyclone activity, there are other important factors as well. Even if sea surface temperatures are high—a favorable factor for development—other factors may mitigate tropical storm formation. High amounts of vertical wind shear, for example, are unfavorable for development. Such conditions typically prevail over the Caribbean during El Niño years, which are consequently unfavorable for Atlantic TC activity. La Niña years, by contrast, are favorable for Atlantic TC activity. Another climate pattern known as the North Atlantic Oscillation [65] ('NAO'), which represents the year-to-year changes in the configuration of the jet stream over the North Atlantic, also has an influence on Atlantic TC activity. During years with a tendency for the negative phase of the NAO, when the Bermuda subtropical high pressure system is weaker than normal, storms are more likely to track through the tropical Atlantic and Caribbean where they encounter more favorable conditions for development. During positive NAO years, storms are more likely to track northward over colder extratropical Atlantic waters.
It is thus possible to relate year-to-year changes in Atlantic TC counts to three basic climate factors: (1) tropical Atlantic sea surface temperatures (SSTs) over the MDR during the August-October season, (2) ENSO, and (3) NAO. The long-term history of annual Atlantic TC counts, along with the time histories of factors (1)-(3) are shown below (for *a*, the red coloring indicates years with greater than average annual TC counts; for *b-d*, the red coloring indicates that the factor in question is more favorable than average for Atlantic TC activity. The Niño3.4 index is a time series that measures the state of ENSO where positive values indicate El Niño events and negative values indicate La Niña events.

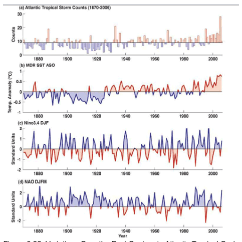


Figure 3.22: Variations Over the Past Century in Atlantic Tropical Cyclone Counts and Various Climate Factors.
From: Mann, M.E., Sabbatelli, T.A., Neu, U., Evidence for a Modest Undercount Bias in Early Historical Atlantic Tropical Cyclone Counts, *Geophys. Res. Lett.*, 34, L22707, doi:10.1029/2007GL031781, 2007.

There are a number of observations that can be made here. First of all, it is clear that annual TC counts have increased over time. How *much* we can conclude they have increased depends on the assumptions regarding how many TCs might have been missed in earlier decades, but even allowing for the upper-end estimates of undercounted past activity, the levels of activity over the past 15 years are without any modern-day precedent. It is also clear that this increase is coincident with increasing sea surface temperatures in the MDR. (Some researchers have argued that much of the long-term variability in temperatures is associated with the AMO mode discussed in the previous section; others, however, have argued that the long-term trends are mostly forced by a combination of human influences including greenhouse gases and sulphate aerosols [66]. If you're interested in Michael Mann's work in this area, you can refer to the publication in the journal *Eos* entitled *Atlantic Hurricane Trends Linked to Climate Change* [67]).

In the absence of other factors, we might assume that any continued warming of the tropical Atlantic in the future will lead to further increases. Yet, as alluded to earlier, there may be mitigating influences as well. We can, for example, see the mitigating impacts on annual TC counts of individual El Niño events if we look carefully at the respective time series shown above. Is it possible that a future trend towards a more El Niño-like climate could offset the effects of warming temperatures on Atlantic TC activity? We will revisit such questions in a later lecture on projected climate change impacts.

The historical relationship between annual TC counts and three climate factors identified above provide us with an ideal application of a new, somewhat more sophisticated statistical tool in our arsenal of methods for analyzing data. We will investigate the method of *multivariate linear regression*, a generalization of the OLS method investigated in our first problem set, which allows us to investigate a problem where there is more than one factor or 'predictor' (in our case, multiple climate factors such as MDR SST, El Niño, and the NAO) that appear to influence some target variable (in this case, annual TC counts) of interest. [*note*: Technically speaking, ordinary linear squares methods, including multivariate linear regression, are not strictly valid for discrete data, i.e., data such as storm counts that take on integer values, i.e., 0, 1, 2...rather than like, say, temperature, which takes on continuous values which better conform to a Gaussian distribution, as discussed in our previous lesson. Strictly speaking, a more sophisticated regression approach known as *Poisson Regression*, is best used in such cases. However, it turns out that, as the counts become large, the assumption of continuous Gaussian behavior becomes increasingly valid. If we were looking at, say, the number of U.S. land falling major hurricanes, where numbers are quite small (typically at most a couple in any given year), that condition would not be met. But for looking at total named TCs, where the average value is about 10 per year, the Gaussian assumption is not too bad, and we can get away with applying standard regression methods.]

You will perform a multivariate regression using these data in your next problem set, [68] using the *multivariate* option in the online Regression Tool [69] that you used in your previous problem set. In the meantime, however, we will use another example to explore how multivariate linear regression works...

Multivariate Regression Demonstration

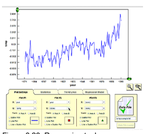


Figure 3.23: Regression tool.

We will now investigate the multivariate generalization of ordinary linear regression, using a data set of Northern Hemisphere land temperature data over the past century. We will attempt to statistically model the observed data in terms of a set of three predictors: (1) estimates from a simple climate model (discussed in our next lesson) known as an Energy Balance Model [70] that has been driven by estimated historical anthropogenic (greenhouse gas and aerosol) and natural (volcanic and solar) radiative forcing histories, and two internal climate phenomena discussed in the previous subsection: the (2) DJF Niño3.4 index, measuring the influence of the El Niño phenomenon, and the (3) DJFM NAO index.

The demonstration is in 4 parts below (click each link to open a new window and then the arrow to begin the demonstration):

Part 1 [71]

Part 2 [72]

Part 3 [73]

Part 4 [74]

You can play around with the data sets used in this example yourself using the Linear Regression Tool. [75]

Extreme Weather

We have already looked at the relationship between climate (and climate change), and one particular type of severe weather—tropical cyclones & hurricanes. Let's now consider other types of severe weather and possible connections with climate change.

Heat Waves

It is perhaps obvious that global warming leads to more frequent and intense heat waves. What is not so obvious, however, is just how profound an impact even modest warming has on the frequency of heat extremes. It all has to do with the statistical properties of our friend, the Gaussian distribution [76].

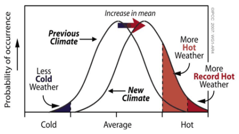


Figure 3.24: Demonstration of How A Shift in Average Temperature Influences the Frequency of Extreme Heat.
Credit: IPCC 4th Assessment Report, Working Group 1 Report

Note from this schematic that even a modest warming can lead to a dramatic increase in the shaded region that exceeds some threshold (i.e., that exceeds 1 or even 2 standard deviations above the mean). Let's work out a simple example based on things we already know about the standard deviation. Let us consider a hypothetical city where the mean daily high temperature in July is 28°C (82°F), and the standard deviation in that measure (i.e., the amplitude of typical day-to-day fluctuations) is 5°C (9°F). Then, using what we know about the standard deviation, roughly 16% of the time, the daily high temperature would be expected to exceed 1 s.d. above the mean (i.e., 33°C/91°F). [You can check this out yourself using Vassar's online calculator [77], setting z = 1 and using the one-sided test result.] Only roughly 2.5% of the time would it be expected to exceed 2 s.d. above the mean (i.e., 38°C/100°F). [Again, you can check this with the online calculator, using z = 2 and observing the one-sided result.] Put in even more basic terms, we would only expect one day in July when the temperature would exceed the 'century mark' of 100°F.

Now, consider the effect of a hypothetical warming of 1°C/2°F (the rough actual warming of the globe since pre-industrial time). We can represent the effect of this warming by shifting the entire temperature distribution to the right by 1°C as shown qualitatively in the schematic above. Now, with the mean of 29°C (and assuming the standard deviation is still 5°C), a 'century mark' temperature of 38°C/100°F is only 1.8 s.d. above the mean.

Think About It!

Use Vassar's online calculator [77] to determine the probability of exceeding the 'century mark' after a warming of 1°C/2°F has occurred.

Click for answer.

So, is there evidence that this is really happening? Indeed, there is. Let's start with the single example of the 2003 European heat wave. This wasn't just 'another heat wave'. It killed more than 30,000 people as a result of exposure to extremely high temperatures coupled with a lack of widespread access to modern air conditioning in large parts of Europe. The entire summer was unusually warm, making individual record breaking heat waves exceptionally more likely.

In Europe, there are reasonably reliable temperature records stretching back several centuries (including thermometer measurements back to the late 18th century—see figure below, and longer-term documentary evidence stretching back as far as 500 years). The 2003 summer temperatures were by far the warmest on record. So, there was a context for the summer 2003 European heat wave. It wasn't just a single random event, but it occurred in the larger-scale context of an unusually warm summer, imbedded in a long-term trend of warming European summer temperatures. One prominent study [78] in the journal Nature in 2004 suggested that global warming had already played a significant role in the European heat waves, taking what might have been thought as a one-in-a-thousand-year event (what is termed a 'thousand year event') and instead turning it into a 20 year event. Additional projected warming, the authors argued, would turn it into a 2 year event, i.e., every other summer would have similar heat waves.

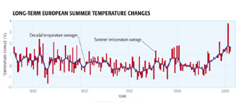


Figure 3.25: Long-term European Summer Temperature Changes.
Credit: Mann & Kump, *Dire Predictions*

That unusually warm summer was associated with a poleward expansion of the jet stream relative to its typical position and a migration of the warm, dry descending air usually found in the descending limb of the Hadley circulation that is typically located in the subtropics (e.g., the Sahara desert), well into Northern Europe (see figure below).

Figure 3.26: Subtropical Zone Expansion.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Indeed, as we will see later in the course when we cover projected changes in atmospheric circulation [32], this pattern of a poleward migration of the descending limb of the Hadley circulation is a robust prediction of state-of-the-art climate models. Viewed in this context, the 2003 European summer heat wave is probably as good an example as any of the potential impact of anthropogenic climate change on heat extremes.

The year 2010 saw record heat around the globe. Perhaps best publicized was the record-breaking heat wave in Moscow and western Russia [79], where new all-time temperature records (111°F) were set, and temperatures hovered around or above 100°F for most of July, replete with massive wildfires and dangerously poor air quality. However, a large number of other countries set new heat records for maximum warmth, including Finland, Pakistan, Sudan, Saudi Arabia, Iraq, and Pakistan.

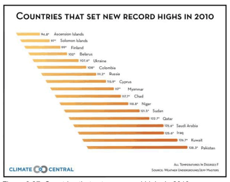

Figure 3.27: Countries that set new record highs in 2010.
Credit: Climate Central [80]

One might rightfully argue that pointing to any one year, be it 2003 in Europe, or 2010 in many other countries, is cherry-picking. And indeed, we need to look at the broader picture in which this fits.

The plot below shows the change over the course of the latter half of the 20th century in the number of days per decade qualifying as unusually warm (defined as exceeding the 90 percentile). Most, though not all, regions have seen an increase in the frequency of extremely warm *days*. Even more striking is the fact that virtually all regions have seen increases in the frequency of extremely warm *nights*. As we will discuss further in our assessment of climate change impacts [81] later in the course, it is actually the latter feature—the increase in very warm nights—that represents a greater threat from the standpoint of human mortality.

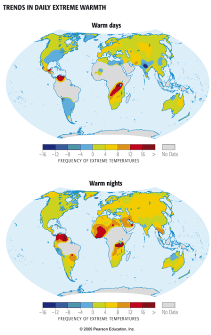

Figure 3.28: Trends in daily extreme warmth.
Credit: Mann & Kump, *Dire Predictions*

What about closer to home, i.e., the U.S.? 2010 was a record-breaking year for heat in many respects for the U.S.

On any given day, just by chance alone, there are likely some places in the U.S. where it is unusually cold, and other places where it is unusually warm. Record cold and record warm are often likely to be found somewhere. The real question to ask is, if we look at all of the reporting locations in the U.S. over all of the days of year, how often are we breaking warm records vs. cold records. As temperatures warm overall, we expect cold records to increasingly be outstripped by warm records. This was certainly the case for 2010.

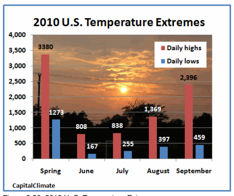

Figure 3.29: 2010 U. S. Temperature Extremes.
Credit: Capital Climate

Not surprisingly, summer (June-August) 2010 was the warmest, or one of the few warmest, summers ever for a large swath of the southeastern and mid-Atlantic U.S. [for example, in State College, PA, daily maximum temperatures in the summer exceeded 90°F more than twice as often as usual] and were unusually warm over much of the rest of the U.S. Only in the northwestern corner of the U.S. was it relatively cool.

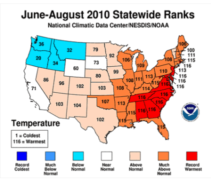

Figure 3.30: June - August 2010 Statewide Ranks.
Credit: NOAA [82]

Again, one might be tempted to argue that focusing on any particular year, including this past year, is cherry-picking. So let us again step back, and look at the long-term context for this anomalous recent warmth. We will look at the trend from decade to decade in the frequency of warm and cold record-breakers, totaled over all locations of the U.S. and over all days of the year. This is what the trend looks like:

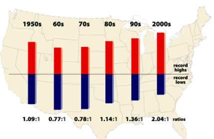

Figure 3.31: Change Over Time in the Relative Incidence of Warm and Cold Daily Extremes.
Credit: UCAR [83]

In the absence of global warming, one would expect record highs to occur as often as lows. This was true in the middle of the century. Then during the 60s and 70s as mean northern hemisphere temperatures—as we have seen before—actually cooled a bit, cold records began to slightly outpace warm records. Since then, we have seen a dramatic increase in the ratio of warm to cold record-breakers as the globe, northern hemisphere, and North America have continued to warm. This trend culminates in a warm/cold record-breaker ratio of more than two-to-one for the past decade. In other words, we've now seen a *doubling* in the relative frequency of hot vs. cold temperature extremes. Note that this increase (roughly a factor of two) in probability of heat extremes is similar to that motivated by the simple example I used at the very beginning of this section on heat waves.

An excellent analogy for the impact of global warming on temperature extremes involves the simple rolling of a six-sided die. So we're going to do a little experiment with die rolling—I think you'll find it both fun and instructive (and amusing—a special thanks to Penn State's David Babb for his

help in constructing this!).

Start rolling the pair of dice. One of the dice is a fair die, and the other is "loaded", though just how, you'll need to figure out. It should be come clearer and clearer as time goes on, and the number of rolls increases. Note that you can "roll" in rapid succession to get larger and larger samples (you don't need to wait for the animation to complete on each roll). Start out with 1, then 5, then 10, 30, 50, 100, and so on, as many as 500 or more if you have the patience! rolls of the die. Record the number of sixes you've rolled for each of the two dice (the % of time each possible value of the die is rolled is conveniently indicated by the changing bar graph off to the right).

Do your rolls seem to be converging towards some well-defined fraction? What is that fraction of the time you are rolling sixes for each of the two dice? How does it compare with your expectation for a *fair* die? When you think you're ready to guess which of the two dice is loaded, go for it. You can repeat the experiment over and over again. Sometimes it's the red die that will be loaded, other times it will be the blue die. How quickly can you successfully identify which is which?

Think About It!

Can you figure out *how* I loaded the die?

Click for answer.

So, as you've figured out by now, I loaded the die so that sixes would come up twice as often as they ought to. The more rolls of the die you do, the more obvious that becomes. Consider your first incidence of rolling a six with the loaded die. Was the fact that you rolled a six on *that roll* directly attributable to the loading of the die? Of course not, you might rightly respond! Even with a fair die, there was a chance (1 in 6) of rolling a six. But the loading of the die nonetheless made it twice as likely that, on any given roll—including that one—you would roll a six. And this becomes increasingly clear as you roll the die more times.

This is a very useful analogy for talking about the influence of global warming on weather extremes such as heat waves. In our first example, we showed that a moderate amount of warming—equivalent to that which has occurred on average over the past century—nearly doubled the probability of breaching the 'century mark' of 100°F during mid-summer. We furthermore saw above that, on average, the relative frequency of extremely warm days in the U.S. has roughly doubled since the mid 20th century. Using the die rolling analogy, we can think of these changes as the equivalent of loading the die in the way we did in the above example. And just as we can't say for certain that any one extremely hot summer day was due to global warming, we can say that the chances were twice as high—and because of global warming. We are indeed seeing the loading of the weather die in the trends toward greater heat extremes in the U.S., and elsewhere.

Other Weather Extremes

Climate change appears to have influenced other types of meteorological extremes (see table below). Not surprisingly, for example, extremely cold days, early frosts, etc., have decreased. Extremely heavy precipitation events and flooding episodes appear to have increased, consistent with the more vigorous hydrological cycle expected with a warming atmosphere. Extratropical cyclones (i.e., mid-latitude storm systems) appear to have strengthened, though the confidence on this observation is somewhat less. Ironically, on the day the first version of this lecture was written (October 27, 2010), the mid-west of the U.S. had just experienced the strongest extratropical cyclone on record [84], with a central low pressure of 954 millibars (that is lower than that of many category 2 Hurricanes!). This is what the October 2010 'superstorm' looked like:

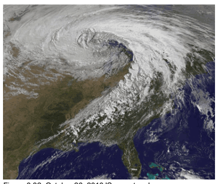

Figure 3.32: October 26, 2010 'Super-storm'.
Credit: NASA

Modeling studies suggest that human-caused climate change may actually decrease the number of extratropical cyclones because the projected polar amplification of warming is likely to reduce the equator-to-pole temperature gradient, which is ultimately what drives these storms through the process of baroclinic instability [57]. However, it may at the same time increase the intensity of the storms systems that do form, because the greater amounts of water vapor available in a warmer atmosphere, once condensed into precipitation as the air rises along frontal boundaries, yield additional latent heating that can add to the energetics of the storm. You can find an excellent discussion of this topic here [85] on Jeff Master's Weather Underground site).

Table 3.1: Table Documenting Potential Climate Change Impacts on Various Types of Extreme Weather.
Credit: IPCC 4th Assessment Report, Working Group 1 report, Chapter 3, Table 3.8

Phenomenon	Change	Region	Period	Confidence	Section
Low-temperature days/nights and frost days	Decrease, more so for nights than days	Over 70% of global land area	1951 - 2003 (last 150 years for Europe and China)	Very likely	3.8.2.1
High-temperature days/nights	Increase, more so for nights than days	Over 70% of global land area	1951- 2003	Very likely	3.8.2.1
Cold spells/snaps (episodes of several days)	Insufficient studies, but daily temperature changes imply a decrease				
Warm spells (heat waves) (episodes of several days)	Increase: implicit evidence from changes in inter-seasonal variability	Global	1951 - 2003	Likely	FAQ 3.3
Cold seasons/warm seasons (seasonal averages)	Some new evidence for changes in inter-seasonal variability	Central Europe	1961 - 2004	Likely	3.8.2.1
Heavy precipitation events (that occur every year)	Increase, generally beyond that expected from changes in the mean (disproportionate)	Many mid-latitude regions (even where reduction in total precipitation)	1951 - 2003	Likely	3.8.2.2
Rare precipitation events (with return periods > ~10 yrs)	Increase	Only a few regions have sufficient data for reliable trends (e.g. UK and USA)	Various since 1893	Likely (consistent with changes inferred for more robust statistics)	3.8.2.2
Drought (season/year)	Increase in total area affected	Many land regions in the world	Since 1970's	Likely	3.3.4 and FAQ 3.3
Tropical cyclones	Trends towards longer lifetimes and greater storm intensity, but no trend in frequency	Tropics	Since 1970's	Likely; more confidence in frequency and intensity	3.8.3 and FAQ 3.3
Extreme extratropical storms	Net increase in frequency/intensity and poleward shift in track	NH land	Since about 1960	Likely	3.8.4, 3.5 and FAQ 3.3
Small-scale severe weather phenomena	Insufficient studies for assessment				

Paleoclimate Evidence

Among contrarians in the public debate about climate change, one often hears an argument that goes something like this:

"*Sure, the climate is changing. But climate is always changing. It was warmer than today in the past due to natural causes! So the warmth today could also be due to natural causes?*"

Is this a legitimate argument? Well, we began to address the issue back during our introduction to the concept of climate change [86]. Let's explore the issue in more detail now, looking at what observations are available to document (a) how climate changed in the past and (b) what factors appear to have been responsible for those changes.

Geological Variations

Let's start out with a look at the longest timescale changes that are documented in the geological record. We're talking hundreds of millions of year timescales. While the records are imperfect at best, we do have a rough idea from very old sedimentary records as to the broad-scale changes in global temperatures and in atmospheric CO_2 over these timescales. On these very long timescales, the changes in atmospheric CO_2 are related to geologic processes such as plate tectonics, which govern the outgassing of CO_2 from Earth's interior by volcanoes, and the slow changes in Earth's continental configuration and relief, which influence natural processes such as chemical and physical weathering that take CO_2 out of the atmosphere.

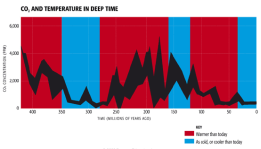

Figure 3.33: CO_2 and Temperature in Deep Time.
Credit: Mann & Kump, *Dire Predictions*

Its quite clear from the comparison above that, on these long timescales, atmospheric CO_2 levels and global temperatures appear to be very closely correlated. Warm periods, with some exceptions, are periods of high atmospheric CO_2, and cold periods, geologically, have been periods of low atmospheric CO_2. You can hear Penn State professor Richard Alley's eloquent Bjerknes Lecture [87] on the subject at the 2009 meeting of the American Geophysical Union.

Let's zoom in further over the past 50 million years.

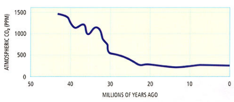

Figure 3.34: Geologic Records of CO_2 Levels.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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This interval takes us from the warm early Eocene period [88], to the cooler Miocene period [89], and eventually to the Plio-Pleistocene period [90] of the past 5 million years, during which large-scale glaciation of the modern continents was first observed. We see that this transition from a warm to glacial climate was accompanied by a major decrease in atmospheric CO_2 levels.

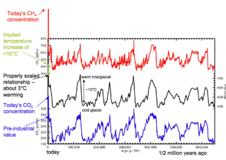
The decrease in atmospheric CO_2, once again, doesn't explain *all* of the details of course. For example, CO_2 appears relatively flat for the final two million years, even as the climate continues to have cooled through the Plio-Pleistocene transition. Some of that cooling may represent positive ice albedo feedbacks that kick in once large-scale glacial inception takes hold over the past 20-30 million years. Nonetheless, CO_2 clearly remains the major knob controlling global climate over the past 50 million years. Once we zoom in on yet shorter timescales, that is tens of thousands of years, we see that another dynamic appears to have been at play.

Glacial/Interglacial Variations

Once we enter into the Pleistocene period of the past 1.5 million years, large-scale glaciation of the Northern Hemisphere takes hold, but there are prominent alternations between relatively ice-covered periods (the "Ice Ages"), and warmer periods more similar to modern, pre-industrial conditions. It is instructive to consider what factors appear to have been driving these variations.

Pleistocene Ice Ages

For nearly the past million years, CO_2 concentrations, as recorded in Antarctic ice cores, have oscillated by roughly 100 ppm, alternating between low *glacial* levels of roughly 180ppm and relatively high *interglacial* levels of roughly 280 ppm. Methane (CH_4) concentrations oscillate by nearly a factor of two as well. And temperatures in Antarctica, as recorded by oxygen isotopes in the ice cores, have varied in concert with these greenhouse gas concentrations changes. If one takes into account the typical polar amplification [91] of warming, the peak-to-peak changes in temperature in Antarctica (roughly 10°C) translate to a change in global mean temperature of about 3°C. At least half of that temperature change is due to changes in Earth's *albedo* (reflectivity) associated with the changes in surface ice cover. That leaves the remaining 1.5°C temperature change to be explained by the changes in greenhouse gas concentrations (CO_2 and also CH_4). If we work out the implied sensitivity of the global climate to greenhouse gas concentrations, it implies a value that lies within the typical range of estimates [66], i.e., between 1.5° and 4.5°C for a doubling of CO_2 concentrations. (You can read more about the details of this argument in *The lag between temperature and CO_2* [92] article from Real Climate).



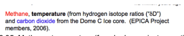
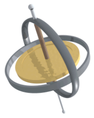


Figure 3.35: Methane, temperature (from hydrogen isotope rations ("δD"), and carbon dioxide from the Dome C ice core.
Credit: EPICA [93]

But why do the oscillations occur? Why do they have a roughly 100,000 year ('100 kyr') periodicity? And why is the shape of the oscillation not a sinusoidal cycle, but a "sawtooth" waveform with a slow, long-term descent into glacial conditions and a very rapid termination? These features certainly require further explanation.

The pacing of the 100 kyr oscillation is almost certainly tied to long-term changes in *Earth orbital geometry*. As I alluded to in the introductory lecture of the course [94], the geometry of Earth's annual orbit around the Sun changes slowly over time. These changes are subtle, but they are persistent over thousands of years and have a profound impact on climate.

The first of these orbital variations involves the very slow wobble (or to use the more technical term, the precession [95]) of the Earth's rotational axis. One full wobble (analogous to the wobbling of a gyroscope—see animation below) takes roughly 19-23 kyr. The precession determines when the Northern and Southern hemisphere are each tilted toward (summer) or away (winter) from the Sun.



Gyroscope.
Credit: Wikimedia Commons [96]

The primary importance of this factor is that it determines whether the summer solstice in a given hemisphere occurs when Earth is farthest (making summer a little cooler) or closest (making summer a little warmer) to the Sun. This factor only matters, then, because Earth's annual orbit around the Sun is not circular, but slightly elliptical - a factor discussed further below.

The second of these changes involves the tilt angle (or to use the more technical term, the obliquity [97]) of Earth's orbit. The Earth's rotational axis today is inclined at an angle of roughly 23.5 degrees from the vertical (this is the angle of the precessing top shown in the animation above). This is why the tropics are located at 23.5N and 23.5S and why the Arctic and Antarctic circles are located at 90-23.5 = 66.5 N and S respectively. This angle of inclination is not fixed over time, however. It varies between roughly 21.5 degrees and 25.5 degrees. Seasonality only exists because of the tilt; if not for the tilt, neither hemisphere would be preferentially tilted towards the Sun at any time of the year. Therefore, periods when the tilt angle is greatest are periods of heightened seasonality, while periods when the tilt angle is smallest have reduced seasonality. It takes roughly 41 kyr for the tilt angle to go through one full cycle of alternation between lesser and greater values of the obliquity.

The final of the changes involves the ellipticity (or, to use the more technical term, the eccentricity [98]) of the orbit. Earth's orbit around the Sun is not circular, but, instead, is *slightly* elliptical. The degree of ellipticity is measured by the *eccentricity*, which ranges from roughly zero (an essentially circular orbit) to a maximum of roughly 4% (a very slightly elliptical orbit). It takes roughly 100 kyr for the eccentricity to go through one full cycle of alternation between low and high eccentricity.

This might sound like the smoking gun to explain the dominant 100 kyr periodicity of the ice ages. But it's not quite that simple.

The changes in insolation associated with the 100 kyr eccentricity variations is considerably smaller than the shorter (19-23 kyr and 41 kyr) precession and obliquity periodicities. And prior to about 700,000 years ago, the glacial/interglacial cycles were dominated by these shorter periodicities. So what is it that is responsible for the dominant 100 kyr oscillation of the past 700,000 years?

To understand this, we have to think a bit more about how these various effects interact. The precession and obliquity don't actually change the net solar radiation at the top of the atmosphere; they simply redistribute it with respect to season and latitude. The eccentricity, however, does change the net solar radiation. When Earth's orbit is more elliptical, it spends more time over the course of a year being relatively far away from the sun, and thus there is a decrease in the received solar radiation. Because the change is very small, however, it's necessary to have large amplifying feedbacks to give this effect a greater role.

Think About It!

Any guess as to *what feedbacks* may do the trick?

Click for answer.

The ice-albedo feedback is key to understanding the glacial/interglacial cycles. While this feedback is a moderate player in the context of modern climate change (responsible, for example, for only about 0.6°C of the 2.5-3°C warming [66] expected for a doubling of CO_2 concentrations), it is considerably more important during ice ages, when continental ice sheets reached as far south as New York City.

This observation provides a clue as to how the 100 kyr periodicity emerges when the actual radiative forcing at the 100 kyr timescale is so weak. Roughly every 100 kyr or so, the eccentricity reaches a particularly high value, favoring the largest possible differences in Earth-Sun distance over the course of the annual cycle. During such times, there will be a point over the course of the much faster 19-23 kyr precession cycle, where the Northern Hemisphere winter solstice will coincide with the minimum Earth-Sun distance ('perihelion'), and the summer solstice will coincide with the maximum Earth-Sun distance ('aphelion'). This makes winters at high northern latitudes unusually warm and summers at high northern latitudes unusually cool. Such conditions are ideal for growing ice sheets: the warm winters actually allow for greater amounts of snowfall (since warm winter air holds more moisture than cold winter air), and the cool summers help prevent any accumulated winter snow from melting over the course of the summer. Ice begins to accumulate, more solar radiation is reflected to space, and the cooling spreads southward. Pretty soon, you've got a full fledged ice age on your hands. This effect, of course, is enhanced when the obliquity is greatest.

So, we can see how all three factors—the eccentricity, the precession, and the obliquity—work together to create ice ages. But the thing that appears to trigger it all is the slow, 100 kyr eccentricity forcing, since high eccentricity—reached roughly every 100 kyr—sets up the perfect storm of conditions for establishing an ice age. Once those conditions occur, an ice age slowly sets in and builds, for tens of thousands of years. This is the slow descent into the ice ages seen in the ice core figure earlier. Eventually, however, as we near 100 kyr from the starting point, the eccentricity again rises to high levels. Then, when the faster precession cycle reaches a point where seasonality is enhanced rather than diminished, ie., northern summers are especially warm—the ice begins to melt, and the positive ice albedo feedback kicks in in an especially dramatic fashion, yielding the rapid warming that defines the abrupt termination evident in the ice core figure.

The mechanism described below requires the possibility of building large ice sheets. It is widely speculated that the climate had to cool beyond some threshold where large ice sheets could grow, and that the slow cooling was providing by the gradual lowering of CO_2 concentrations over the course of the Pleistocene. Eventually the climate cooled to that point, and the 100 kyr oscillations ensued.

And what about the role of CO_2 in all of this? As I noted above, CO_2 is clearly an important player. We cannot explain the extent of the warming and cooling over the glacial/interglacial cycles without including the direct warming effect of CO_2 . But the situation is somewhat more complicated than that since CO_2 is not simply a control variable on these timescale. The global carbon cycle (in particular the oceanic carbon cycle) is changing in response to the climate changes taking place. For example, a warming ocean favors outgassing of CO_2 into the atmosphere, which of course amplifies the warming further because of the greenhouse effect.

So CO_2 is acting both as a *forcing* and as a *feedback*. That is different from the current situation, where we essentially have our hands on the CO_2 knob of the system (though, even here, there are some complications, as we will discuss in more detail in Lesson 6 on carbon emission scenarios [99]).

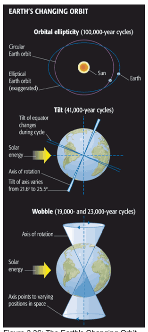


Figure 3.36: The Earth's Changing Orbit.
Credit: Mann & Kump, *Dire Predictions*

The Younger Dryas

At the termination of the last ice age, roughly 12 kyr ago, something surprising happened. Just as Earth appeared to be coming out of the ice age, it staged a reversal and headed back into glacial conditions for at least 1000 years. The cooling event seems to have been centered in the North Atlantic ocean. This episode is known as the *Younger Dryas* (named after the spread of the tundra-loving Dryas flower [100], whose range during this time interval plunged southward in regions surrounding the North Atlantic ocean).

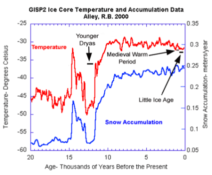


Figure 3.37: GISP2 Ice Core Temperature and Accumulation Data.
Credit: NOAA [101]

While there is still some scientific debate about the details, it is generally believed that this rapid cooling resulted from a slowdown or collapse of the ocean's thermohaline circulation in response to the massive flux of freshwater into the high-latitude North Atlantic as the Northern Hemisphere ice sheets and glaciers rapidly began to melt. This fresh water would have inhibited the sinking motion that typically occurs in the sub-polar regions of the North Atlantic, suppressing the overturning circulation associated with the North Atlantic drift current which helps to warm the high-latitudes of the North Atlantic and surrounding regions.

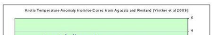
Credit: NASA [102]

It is, in fact, this event from the distant past that has given rise to the popular, if rather flawed (as we will see when we look at actual climate projections [103]), concept that global warming may ironically lead to another ice age. This concept was caricatured in the movie 'The Day After Tomorrow [104]'. The problem with the analogy is that there isn't nearly the amount of ice around today to melt that there was at the termination of the last ice age, and thus the effect is likely to be much smaller. As we will see later [103]in the course, however, there may be a *very* small grain of truth to the scenario laid out in the movie.

The Holocene

Even during interglacial intervals, when there is relatively little ice around, the higher-frequency orbital forcings still have an influence on climate. In particular, the impacts of the roughly 19-23 thousand year precession cycle are evident over the course of the current interglacial period of the past 12,000 years known as the *Holocene*. Today, precession has the effect of *minimizing* Northern Hemisphere seasonality, since perihelion occurs very close (January 3) to the winter solstice (Dec 22nd). Roughly 1/2 of a precession cycle (i.e., about 11,000 years) ago, precession had the effect of *maximizing* Northern Hemisphere seasonality. Indeed, it was this exaggerated seasonality that helped end the last ice age by favoring summer melting. Over the next 12,000 years, the seasonal pattern of insolation slowly evolved to what it is today.

If we look at temperature estimates based on oxygen isotopes from Arctic ice cores, we find that summer temperatures were relatively high during a period sometimes called the *Holocene optimum* that lasted from about 10,000-6,000 years ago, before slowly declining towards pre-industrial levels (and then spiking again over the last 100 years—something we'll discuss more below). This pattern of high-latitude summer temperature change is precisely what we expect, given the precession changes discussed above. On the other hand, tropical insolation was actually reduced, as was winter insolation at higher northern latitudes. That makes these natural past changes in radiative forcing very different from those associated with modern greenhouse increases which are positive during both seasons, in both hemispheres, and in the tropics as well as high latitudes.



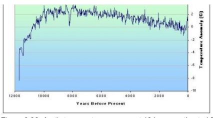


Figure 3.38: Arctic temperatures over past 12 kyr as estimated from ice core oxygen isotopes.
Credit: Skepticalscience.com [105]

As the seasonal and latitudinal pattern of solar heating evolved over the course of the Holocene, so did the atmospheric circulation which is driven in substantial part by those variations in heating. For example, paleoclimate evidence suggests that the Sahara desert was considerably more lush than today, supporting steppe vegetation and large lakes that were home to crocodiles and hippopotamuses. Simulations with climate models, forced by the known changes in seasonal insolation, reproduce this pattern, as a consequence of strengthened west African monsoon, driven by the larger seasonal changes in insolation.

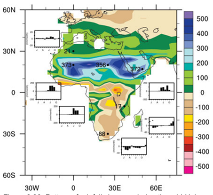


Figure 3.39: Pattern of rainfall changes during the mid-Holocene.
Credit: NESL's Climate & Global Dynamics [106]

The Past 1,000 Years

Over the past 1,000 years, changes in insolation due to Earth orbital effects are relatively small. The primary natural radiative forcings believed to be important on this time frame (as we will see in Lesson 5 when we talk about Estimating Climate Sensitivity [107]) are volcanic eruptions and changes in solar output.

The basic 'boundary conditions' on the climate (distribution of ice sheets, orbital geometry, continental configuration, etc.) were essentially the same as they are today, making this a useful interval to study from the standpoint of modern climate change. The pre-industrial interval of the past 1,000 years forms a sort of 'control', as the only real difference from today is the added impact over the past century or two of human influence on climate. Some (in particular, the atmospheric chemist Paul Crutzen, who won the Nobel prize for his work on the stratosphere ozone hole) have argued that the period since the dawn of industrialization (i.e., the past two centuries) has been impacted substantially enough already by human activity that we should consider it as a distinct interval separate from the current interglacial, i.e., that we are no longer in the Holocene, but rather, an entirely new, human-created interval of Earth history known as the *anthropocene* [108].

Patterns of surface temperature can be estimated in past centuries from "proxy" records such as tree-rings, corals, ice cores, faunal remains and other sources. For more information about the work that Michael Mann has done on this topic, you might check out an Interactive Presentation of the Global Temperature Patterns in Past Centuries [109].

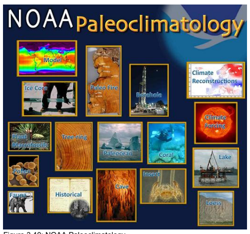


Figure 3.40: NOAA Paleoclimatology.
Credit: NOAA [110]

Looking at the temperature patterns reconstructed from these proxy records, we can gain a longer-term perspective on the natural climate variability associated with ENSO [111], and the NAO [112], and other known dynamical patterns influencing the climate system. Can you spot some big El Niño events in the 18th century? What about the effect of the NAO—can you discern that in the patterns? How?

Credit: NOAA [113]

If we average over these patterns (focusing on the Northern Hemisphere, since information in the Southern Hemisphere is limited) , we can obtain an estimate of the average temperature in past centuries. There are numerous reconstructions that have been performed of this sort, using different types of proxy data, and different statistical approaches to estimating temperatures from the data. But one thing they all have in common is finding that the recent warming is anomalous as far back as these reconstructions go (more than 1,000 years now).

Does this mean that the warming is due to human activity? No—it's possible that the warming could just be a fluke of nature. To assess whether or not the recent warming can be attributed to human activity, we'll need to turn to *theoretical climate models* - the topic of our next lesson!

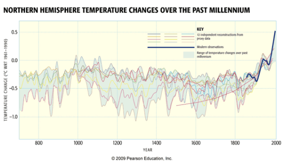


Figure 3.41: Northern Hemisphere Temperature Changes Over the Past Millennium.
Credit: Mann & Kump, *Dire Predictions*

Problem Set #2

Activity: Statistical Analysis of Atlantic Tropical Cyclones and Underlying Climate Influences

NOTE: For this assignment, you will need to record your work on a word processing document. Your work must be submitted in Word (.doc or .docx) or PDF (.pdf) formats.

For this activity, you will perform an analysis of the relationship between Atlantic Tropical Cyclone (TC) counts and three potential climate-related predictors of TC activity, over the time interval 1870-2006.

Link to Linear Regression Tool [69]

Directions

1. First, save the Problem Set #2 Worksheet [114] to your computer. You will use this word processing document to electronically record your work in the remaining steps.
   - Save the worksheet to your computer by right-clicking on the link above and selecting "Save link as..."
   - The worksheet is in Microsoft Word format. You can use either Word or Google Docs (free) to work on this assignment. You will submit your worksheet at the end of the activity, so it must be in Word (.doc or .docx) or PDF (.pdf) format so the instructor can open it.
   - Please show your work! When you are explicitly asked to create plots in a question, please cut-and-paste graphics and the output from the screen (e.g., by first printing the output to a pdf file and then directly inserting into the worksheet) to submit along with your discussion and conclusions.
2. We will start out using the "unadjusted"□ Tropical Cyclone (TC) counts, i.e., counts not adjusted for potential historical undercount of storms in earlier decades. Please separately perform three simple ordinary linear regressions for TC counts (Target Observation = TC_no_adjust) using the following predictors: (A) sea surface temperature (SST) in the main development region (MDR) (Model Parameters = mdr), (B) Niño 3.4 (Model Parameters = niño), and (C) NAO (Model Parameters = nao). For each of the three regressions answer the following questions.
   1. What is the correlation coefficient? Is the relationship (i.e., the sign of the correlation coefficient) consistent with your expectations based on the earlier readings in the chapter regarding the factors that influence Atlantic TC counts?
   2. Is the correlation statistically significant?
   3. Does autocorrelation of the residuals appear to be a problem? Do you see any evidence of additional structure (i.e., trends) in the residuals? What can you conclude about the robustness of this linear regression?
3. For each of the three simple ordinary linear regressions you performed in Question 2, answer the following:
   1. How much variation in the TC count data does the predictor variable account for? Recall that R^2 is a measure of the fraction of variation in the data that is captured by the predictor variable.
   2. Does the predictor capture the long-term upward trend in TC counts? Does it capture the year-to-year variability (finer up and down fluctuations from year to year) in TC counts? Does it capture multi-decadal variability (broader up and down fluctuations over several decades)?
4. Next please perform a multivariate regression using all three predictors simultaneously (Target Observation = TC_no_adjust; Model Parameters = mdr, niño, and nao).
   1. How much of the total variation in the data does your regression model capture?
   2. Does autocorrelation of the residuals appear to be a problem?
   3. Briefly discuss why multivariate regression model is superior to the single predictor regressions from Question 2.
5. Repeat the analyses (i) and (ii) in Question 4 using (A) "lightly adjusted"□ TC count series (Target Observation = TC_light_adjust), where a modest undercount bias is assumed prior to aircraft reconnaissance (i.e., and undercount of roughly 1 storm per year before 1944) and (B) "heavily adjusted"□ TC count series (Target Observation = TC_heavy_adjust), where a more substantial underestimation bias is assumed (i.e., roughly 3 storms per year prior to 1944). How, if at all, did your conclusions change?
6. Using the regression model based on the "lightly adjusted"□ TC series, please make a prediction for the number of TCs that we should have seen during the 2016 Atlantic Hurricane season. Perform the following steps:
   1. Assume that:
      - MDR SSTs were roughly equal to their 2005 value: read the value off the SST plot;
      - there was a modest La Niña event: take Niño 3.4 index to be -0.5;
      - NAO was neutral: use the long-term average of the NAO series.
   2. Write down the multivariate equation. Substitute the values of mdr, nino, and nao found in (i) into the multivariate regression equation to obtain your prediction for TC count. Round off your estimate to the nearest whole number and call the result N. Calculate the 95% confidence interval for your prediction, assuming that the interval is N plus or minus 2 x sqrt(N).
   3. Compare your prediction to the prediction made prior to the beginning of the 2016 Atlantic Hurricane season by the Penn State statistical model [115]. The Penn State model predicted 18.9 plus or minus 4.4 storms, which corresponds to between 14 and 24 storms, with the best estimate of 19 storms. Is your prediction consistent with the Penn State statistical model, within the statistical uncertainty?
   4. Compare your prediction to the actual observed total for 2016: 15 named storms. Is your prediction consistent with the observed counts, within the statistical uncertainty?
7. Save your word processing document as either a Microsoft Word or PDF file in the following format: PS2_AccessAccountID_LastName.doc (or .pdf). For example, student Elvis Aaron Presley's file would be named "PS2_eap1_presley.doc". This naming convention is important, as it will help the instructor match each submission up with the right student!

Submitting your work

Upload your file to the "Problem Set #2" assignment in Canvas by the due date indicated in the Syllabus.

Grading rubric

The instructor will use the general grading rubric for problem sets [116] to grade this activity.

Lesson 3 Summary

In this lesson, we reviewed key observations that detail how our atmosphere and climate are changing. We have seen that:

- ice in its various forms (sea ice, glaciers, and ice sheets) is disappearing as the globe warms;
- global sea level is rising due both to the expansion of warming sea water and the contribution from melting glaciers and ice sheets;
- the deep ocean, as well as the ocean surface, is warming in a manner consistent with a warming climate; the slow nature of the ocean warming means that global sea level will continue to rise for several centuries, even if we were to freeze greenhouse concentrations at current levels;
- ocean circulation trends are difficult to establish, but there is some reason to believe that the ocean's thermohaline circulation may weaken;
- there is substantial natural multidecadal variability that appears related to oscillations in the strength of the ocean's thermohaline circulation;
- the El Niño/Southern Oscillation (ENSO) is the most prominent mode of climate variability on interannual timescales, and one important uncertainty in projecting future climate change involves uncertainties in how ENSO will change in the future;
- there is a trend toward increasingly powerful Atlantic Hurricanes over the past half century, and this increase appears to mirror warming tropical Atlantic sea surface temperatures;
- there is also a trend toward increased Atlantic tropical cyclone counts, though the reliability of the data, particularly in earlier decades, has been called into question. A variety of factors, including sea surface temperatures, El Niño, and the so-called North Atlantic Oscillation (NAO) atmospheric circulation pattern, influence Atlantic tropical cyclone counts;
- other extreme weather events appear to have become more common in recent decades. The increased frequency and duration of heat waves around the world is likely related to large-scale warming; there is also some evidence that mid-latitude storm systems have become more powerful, consistent with predictions of climate models;
- paleoclimate evidence demonstrates that CO_2 has been the 'major lever' influencing global climate change over geological time;
- CO_2 and CH_4 greenhouse gas forcing is a key component in the glacial/interglacial cycles of the late Pleistocene, with ice albedo feedback playing an especially important role, and with changes in Earth orbital geometry providing the pacing, including the dominant 100 kyr oscillations of the past 700,000 years;
- the Younger Dryas cooling, even in the North Atlantic, that occurred at the termination of the last ice age demonstrates the potentially important role of ocean dynamical responses to meltwater fluxes; such effects are unlikely to be as important in the context of modern climate change;
- Earth orbital changes have influenced climate changes over the course of the current interglacial period (the *Holocene*). Increased high-latitude summer insolation was responsible for the relatively warm summers at high Northern latitudes during a period from roughly 10,000-6,000 years ago known as the *Holocene optimum*. The same seasonal and latitudinal redistribution of solar insolation was responsible for changes in atmospheric circulation such as the strengthened west African Monsoon that led to the greening of the Sahara at this time;
- temperatures over the past millennium have been dominated by other radiative forcings, such as natural forcing by volcanic eruptions and solar insolation changes, and anthropogenic forcing over the past two centuries;
- the warming of the past few decades appears to exceed the range seen for at least the past millennium.

Reminder - Complete all of the lesson tasks!

You have finished Lesson 3. Double-check the list of requirements on the first page of this lesson to make sure you have completed all of the activities listed there before beginning the next lesson.

Lesson 4 - Modeling of the Climate System, part 1

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

About Lesson 4

We have now seen evidence indicating that the globe is warming, and that there is an array of other internally-consistent changes in the climate system that are associated with that warming. While these changes are suggestive of human-caused climate change, the existence of trends cannot alone be used to draw causal inferences.

That is where theoretical climate models come in. Climate models allow us to test particular hypotheses about climate change. For example, we can interrogate the models with respect to how much warming of the globe we might expect for a given change in greenhouse gas concentrations. In this lesson, we will consider the simpler classes of climate models, and we will engage in hands-on climate modeling activities.

What will we learn in Lesson 4?

By the end of Lesson 4, you should be able to:

- describe the factors that govern Earth's climate system;
- perform basic energy balance computations to estimate the surface temperature of the Earth;
- perform basic energy balance computations to estimate the response of Earth's surface temperature to hypothetical changes in natural and anthropogenic forcing; and
- explain what "equilibrium climate sensitivity" is.

What will be due for Lesson 4?

Please refer to the Syllabus for specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 4. Detailed directions and submission instructions are located within this lesson.

- Problem Set #3: Estimate the warming due to an increase in CO_2
- Read: Dire Predictions, v.2: p. 68-69

Questions?

If you have any questions, please post them to our Questions? discussion forum (not e-mail), located under the Home tab in Canvas. The instructor will check that discussion forum daily to respond. Also, please feel free to post your own responses if you can help with any of the posted questions.

Energy and Radiation Balance

We actually covered this topic back in the introductory lecture (lecture #1), so I'm going to ask you to simply review the Overview of the Climate System (part 2) [94] before continuing on to our initial discussion of climate models...

Simple Climate Models

We will start out our discussion of climate models with the simplest possible conceptual models for modeling Earth's climate. These models include different variants on the so-called *Energy Balance Model*. An Energy Balance Model or 'EBM' does not attempt to resolve the dynamics of the climate system, i.e., large-scale wind and atmospheric circulation systems, ocean currents, convective motions in the atmosphere and ocean, or any number of other basic features of the climate system. Instead, it simply focuses on the energetics and thermodynamics of the climate system.

We will start out discussion of EBMs with the so-called *Zero Dimensional* EBM—the simplest model that can be invoked to explain, for example, the average surface temperature of the Earth. In this very simple model, the Earth is treated as a mathematical point in space—that is to say, there is no explicit accounting for latitude, longitude, or altitude, hence we refer to such a model as 'zero dimensional'. In the zero-dimensional EBM, we solve only for the balance between incoming and outgoing sources of energy and radiation at the surface. We will then build up a little bit more complexity, taking into account the effect of the Earth's atmosphere—in particular, the impact of the atmospheric greenhouse effect—through use of the so-called "gray body" variant of the EBM.

Zero Dimensional EBM

The zero dimensional ('0d') EBM simply models the balance between incoming and outgoing radiation at the Earth's surface. As you'll recall from your review of radiation balance in the previous section, this balance is in reality quite complicated, and we have to make a number of simplifying assumptions if we are to obtain a simple conceptual model that encapsulates the key features.

For those who are looking for more technical background material, see this "Zero-dimensional Energy Balance Model" online primer [117] (NYU Math Department). We will treat the topic at a slightly less technical level than this, but we still have to do a bit of math and physics to be able to understand the underlying assumptions and appreciate this very important tool that is used in climate studies.

We will assume that the amount of short wave radiation absorbed by the Earth is simply $(1-\alpha)\,S/4$, where *S* is the Solar Constant (roughly 1370 W/m^2 but potentially variable over time) and α is the average reflectivity of Earth's surface looking down from space, i.e., the 'planetary albedo', accounting for reflection by clouds and the atmosphere as well as reflective surface of Earth including ice (value of roughly 0.32 but also somewhat variable over time).

We will assume that the outgoing longwave radiation is given simply by treating the Earth as a 'black body' (this is a body that absorbs all radiation incident upon it). The *Stefan-Boltzman law* for black body radiation holds that an object emits radiation in proportion to the 4th power of its temperature, i.e., the flux of heat from the surface is given by

$$F_{bb} = \varepsilon \cdot \sigma \cdot {T_S}^4$$

(1)

where σ is known as the *Stefan-Boltzmann constant*, and has the value $\sigma = 5.67 \text{ x } 10^{-8}(Wm^{-2}K^{-4})$; ε is the *emissivity* of the object (unitless fraction) — a measure of how 'good' a black body the object is over the range of wavelengths in which it is emitting radiation; and T_s (K) is the surface temperature. For the relatively cold Earth, the radiation is primarily emitted in the infrared regime of the electromagnetic spectrum, and the emissivity is very close to one.

We will approximate the surface temperature, T_S, as representing the average 'skin temperature' of an Earth covered with 70% ocean (furthermore, we will treat the ocean as a mixed layer of average 70m depth—this ignores the impacts of heat exchange with the deep ocean, but is not a bad first approximation). We can then approximate the thermodynamic effect of the mixed layer ocean in terms of an *effective heat capacity* of the Earth's (land+ocean) surface, $C = 2.08 \text{ x } 10^8 \; JK^{-1}m^{-2}$. The condition of energy balance can then be described in terms of the thermodynamics, which states that any change in the internal energy per unit area per unit time (<$\Delta F = CdT_s/dt$) must balance the rate of net heating, which is the difference between the incoming shortwave and outgoing longwave radiation. Mathematically, that gives:

$$C\frac{dT_s}{dt} = \frac{(1-\alpha)\,S}{4} - \varepsilon \cdot \sigma \cdot {T_s}^4$$

(2)

Let's suppose that the incoming radiation (the first term on the right hand side) were larger than the outgoing radiation (the second term on the right hand side). Then the entire right-hand side would be positive, which means that the left-hand side, the rate of change of T_s over time, must also be positive. In other words, T_s must be increasing. This, in turn, means that the outgoing radiation must increase, which will eventually bring the two terms on the right hand side into balance. At this point, there is no longer any change of T_s with time, i.e., we achieve an equilibrium.

In equilibrium, the time derivative term is, by definition, zero, and we thus must have equality between the outgoing and incoming radiation, i.e., between the two terms on the right-hand side of equation 1. This yields the purely algebraic expression

$$\varepsilon \cdot \sigma \cdot {T_s}^4 = \frac{S\,(1-\alpha)}{4}$$

(3)

The factor of 1/4 comes from the fact (see Figure 4.1, below) that the Earth is *emitting* radiation over the entirety of its surface area (4πR^2 where *R* is the radius of the earth), but at any given time only *receiving* incoming (solar) radiation over its cross-sectional area, πR^2.

It turns out that since the Earth's surface temperature varies over a relatively small range (less than 30° K) about its mean long-term temperature (in the range of O° C, or 273° K), i.e., it varies only by at most 10% or so, it is valid to approximate the 4th degree term in equation (1) by a linear relationship, i.e.,

$$\varepsilon \cdot \sigma \cdot {T_s}^4 = A + BT_S$$

(4)

A and *B*, thus defined, have the approximate values: $A = 315 \; Wm^{-2}$; $B = 4.6 \; Wm^{-2}K^{-1}$

Such an approximation is often used in atmospheric science and other areas of physics when appropriate, and is called *linearization*.

Using this approximation, we can readily solve for T_S as

$$T_S = \left[\frac{S\,(1-\alpha)}{4} A\right] / B$$

(5)

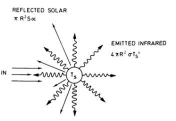
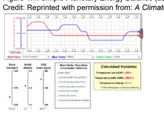


Figure 4.1: Simple Planetary Energy Balance (assume T_e = T_S for our present purposes).
Credit: Reprinted with permission from: *A Climate Modeling Primer*, A. Henderson-Sellers and K. McGuffie, Wiley, pg. 58, (1987).



[118]

We'll now play around a bit with our own customized 0d EBM [118] you'll be using for your problem set. First, we will solve for Earth's mean surface temperature in the *black body approximation*, given reasonable values of the key governing physical parameters (i.e., α and *S*).

Black Body Approximation - Demonstration [119]

You might find it rather disappointing that, after all the work we did above to develop a realistic Energy Balance Model for Earth's climate, we were way off. Our EBM indicates that, given appropriate parameter values (i.e., $S = 1370 \; W/m^2$, $\alpha = 0.32$), the Earth should be a frozen planet with T_S = 255° K, rather than the far more hospitable T_S= 288° K we actually observe. Our model gave a result that was a whopping 33° C (roughly 60° F) too cold!

Think About It!

What do you think we forgot?

Click for answer.

So, how do we include the effect of the atmospheric greenhouse effect in a simple way? That is the topic of our next section.

Simple Climate Models, cont'd

Gray Body Variant of the Zero Dimensional EBM

Even in the presence of the greenhouse effect, the net longwave radiation emitted out to space must balance the incoming absorbed solar radiation. So, we can think of the Earth system as still possessing an *effective radiating temperature* (T_e), which is the *black body* temperature we calculated earlier with the zero-dimensional EBM and the black body parameter values for A and B, i.e., $T = 255^o$ K. It is the temperature Earth's surface has in the absence of any greenhouse effect. The outgoing longwave radiation to space is still given by $\varepsilon \cdot \sigma \cdot {T_e}^4$. The atmosphere will have a temperature T_e somewhere aloft in the cooler region of the mid-troposphere. If we like, we can think of the Earth as, on average, emitting temperature to space from this level; hence, we refer to the temperature as the effective radiating temperature.

When a greenhouse effect is present, the temperature at the surface, T_S, will be substantially higher, however, due to the additional *downward* longwave radiation emitted by the atmosphere back down towards the Earth's surface.

We can attempt to account for this effect by by simply changing the way we model the longwave radiation in the zero-dimensional EBM to account for the additional downward longwave radiation component.

Returning to the linearized form of the energy balance equation (i.e., equation 3 above), we will, therefore, now relax the assumption that A and B are given by their *black body* values. Instead, we will allow A and B to take on arbitrary values. This is a crude way of taking into account the fact that the Earth does not behave as a black body because the atmosphere has non-zero emissivity due to the presence of atmospheric greenhouse gases.

Simply put, we can tweak the values of A and B until they provide a good approximation. We refer to this generalized version of the black body approximation as the *gray body* approximation. The gray body model is a very crude way of accounting for the greenhouse effect in the context of a simple zero-dimensional model. In Lesson 5, we will build our way up to more realistic representations of the atmospheric greenhouse effect.

Various gray body parameter choices for A and B have been used by different researchers, in different situations. Since the gray body approximation is a linear approximation to a non-linear (Planck radiation) relationship, it is only valid over a limited range of temperatures about a given reference temperature. This means that a different set of parameters might be used for studying, e.g., the ice ages than would be used for studying, e.g., the early Cretaceous super greenhouse.

It turns out that the choices $A = 214.4\ W/m^2$ and $B = 1.25\ W/m^2\ K^{-1}$ yield realistic values for both the current average temperature of the earth T_S, and gives a value for the *climate sensitivity* — a concept we will define in the next section — that is consistent with mid-range IPCC estimates. We will, therefore, adopt these as our standard *gray body* parameter values, but we will also explore the impact of using alternative values a bit later.

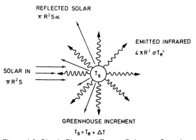


Figure 4.2: Simple Planetary Energy Balance - Greenhouse Effect Included.
Credit: Reprinted with permission from: *A Climate Modeling Primer*, A. Henderson-Sellers and K. McGuffie, Wiley, pg. 58, (1987).

Think About It!

Use the Online 0d EBM Application [118] to estimate the average temperature of the Earth for the "mid-range IPCC" gray body parameter values. What surface temperature do you find, and how does it compare with the previous black body estimate of Earth's surface temperature?

Click for answer.

The Concept of *Equilibrium Climate Sensitivity*

Let us rewrite the equation energy balance equation (3) above in a slightly different form,

$$T_S = \left[F_{in} - A\right]/B$$

(4)

$$T_S = \left({}^1\!/_B\right) F_{in} - {}^A\!/_B$$

(5)

where F_{in} represents the total incoming radiative energy flux at the surface, which includes incoming short wave radiation, but also any potential changes in the downward longwave radiation towards the surface.

Let us now consider the response of T_s to an incremental change in F_{in}. Since the 2nd term in (5) is a constant, we simply have

$$\Delta T_s = \frac{\Delta F_{in}}{B}$$

(6)

We can also rewrite (6) as

$$\frac{\Delta T_s}{\Delta F_{in}} = \frac{1}{B}$$

(7)

The change in downward longwave radiation forcing associated with a change in CO_2 concentration from a reference concentration, $[CO_2]_0$ to some new value, $[CO_2]$, can be approximated by the following relationship from a paper by Myhre et al. (1998) [120]

$$\Delta F_{CO_2} = 5.35 \ln\left(\frac{[CO_2]}{[CO_2]_0}\right)$$

(8)

Now, let us further specify that we are interested in the change in radiative forcing resulting from a *doubling of atmospheric CO_2 concentrations*. For a CO_2 doubling, e.g., an increase from pre-industrial levels of 280 ppm to twice that value, 560 ppm,

$$\Delta F_{2xCO_2} = 5.35 \ln\left(\frac{560ppm}{280ppm}\right) = 3.7\frac{W}{m^2}$$

(9)

We can define *equilibrium climate sensitivity*, *s*, as the change in temperature resulting from a doubling of pre-industrial CO_2 concentrations; *s* has units of K (or equivalently degrees C, since differences in C and K are equal). To estimate s, we combine equations (6) and (9)

$$S = \Delta T_{2xCO_2} = \frac{\Delta F_{2xCO_2}}{B} = \frac{3.7}{B}$$

(10)

The equilibrium climate sensitivity is the equilibrium warming we expect in response to CO_2 doubling. In the simple case of the 0d EBM, it is readily calculated through equation (10).

Think About It!

Using the formula above (10), estimate the equilibrium climate sensitivity *s* for both the *black body* model and our standard version of the *gray body* model. Record your answers.

Click for answer.

Think About It!

Let's now use the Online 0d EBM Application [118] again to estimate the climate sensitivity for these two cases, by explicitly varying the CO_2 level until you achieve a CO_2 doubling, and recording the warming that you observed. Compare to the results you calculated above directly from the formula for climate sensitivity for the 0d EBM.

Click for answer.

Problem Set #3

Estimating Future Warming

Activity

NOTE: For this assignment, you will need to record your work on a word processing document. Your work must be submitted in Word (.doc or .docx) or PDF (.pdf) format so I can open it.

For this activity, you will explore the warming due to increases in CO_2 using a simple (0d EBM) climate model. You will consider applications to:

- defining CO_2 thresholds for avoiding dangerous human impacts on climate;
- the controversial notion of 'geo-engineering' as a means of mitigating human-caused climate change.

Online 0d EBM Application [118]

Directions

1. First, save the Problem Set #3 Worksheet [121] to your computer. You will use this word processing document to electronically record your work in the remaining steps.
    - Save the worksheet to your computer by right-clicking on the link above and selecting "Save link as..."
    - The worksheet is in Microsoft Word format. You can use either Word or Google Docs (free) to work on this assignment. You will submit your worksheet at the end of the activity, so it must be in Word (.doc or .docx) or PDF (.pdf) format so the instructor can open it.
    - Please show your work! When you are explicitly asked to create plots in a question, please cut-and-paste graphics and the output from the screen (e.g., by first printing the output to a pdf file and then directly inserting into the worksheet) to submit along with your discussion and conclusions.
2. Use the online 0d EBM application to calculate the Earth's mean temperature (pre-industrial, that is, without anthropogenic greenhouse warming) and climate sensitivity (the net warming due to CO_2 doubling from pre-industrial) for the (i) low-end, (ii) mid-range, and (iii) high-end IPCC gray body parameter settings. To do that, please use the default settings for the solar constant ($S = 1370\ Wm^{-2}$) and the Earth's albedo ($\alpha = 0.32$), and double the pre-industrial CO_2 concentrations. Discuss how the mean temperature and climate sensitivity change depending on the choice of IPCC gray body parameter.
3. The European Union has defined 2 °C degrees warming relative to the pre-industrial temperatures as the threshold for dangerous anthropogenic interference (DAI) with the climate system. Use the online 0d EBM application to estimate the level of CO_2 at which would we expect to breach the DAI amount of warming for the (i) low-end, (ii) mid-range, and (iii) high-end IPCC gray body parameter settings. Once again, use the default settings for the solar constant and the Earth's albedo. NOTE: that the adjustable slider for CO_2 only goes up to 560 ppm, so to use higher values you need to enter them by hand into the box below the slider. [See here for a commentary on the notion of DAI by your course author [122].]
4. Atmospheric CO_2 is currently at about 400 ppm and is increasing by about 2 ppm per year. If we continue to increase CO_2 at this rate, how many years will it take until we commit ourselves to DAI, based on the three climate sensitivities (i.e., gray body IPCC parameterizations) considered above? To answer this question, use your results from Question 3. If you were advising policy makers, how many years would you tell them we have to stabilize CO_2 emissions and why?
5. Later in this course, we will encounter the concept of geo-engineering — an approach to climate change mitigation which involves the attempt to offset greenhouse warming through various means of intervention with the climate system. One much-discussed geo-engineering scheme involves shooting sulfate aerosols into the stratosphere, mimicking the natural cooling impact of explosive volcanic eruptions. The eruption of Mount Pinatubo in 1991, for example, caused approximately 2.5 W m^{-2} reduction in the amount solar radiation incident at the Earth's surface, which is equivalent to about a 14 W m^{-2} reduction in the solar constant, and the effect lasted over about a 3 year period. (Recall from Lesson 4 that the amount of solar radiation incident at the Earth surface, F_{in}, relates to the solar constant, *S*, as $\mathrm{F_{in}} = (1-\alpha) * S/4$, where α is the Earth's albedo.) Therefore, over a 3 year period, Mt. Pinatubo caused about a 1% reduction in the solar forcing. Use the online 0d EBM application to consider a scenario where we stabilize CO_2 concentrations at double their pre-industrial levels. Assuming the mid-range gray body IPCC parameterization, with what frequency would we have to simulate Mt. Pinatubo-like eruptions to keep global mean temperatures from exceeding the 2 °C DAI threshold? HINT: Start by recording temperature (with AGW) for the default values of the solar constant and the Earth's albedo, and double pre-industrial CO_2 for the mid-range gray body IPCC parameterization. This setup produces a 3 °C increase in temperature over the pre-industrial CO_2 levels. Then, decrease the solar constant until you find that the temperature (with AGW) decreased by 1°C, which results in total 2 °C increase in temperature that we need for DAI (ignore Temperature Change field, as it does not change). Based on the change you applied to the solar constant, approximate the required frequency of Mt. Pinatubo eruptions: one way to think of it is that one Pinatubo every 3 years gives you an average 1% reduction in *S*, so one Pinatubo every 6 years gives an average 0.5% reduction in *S*, one Pinatubo every 9 years gives an average 0.33% reduction in *S*, etc.
6. An alternative geo-engineering approach involves changing the Earth's surface properties by altering the albedo through various schemes, e.g., painting rooftops white, [123] or artificially seeding low clouds [124] over the ocean. By what percent would we need to increase Earth's albedo to avoid exceeding 2 °C warming relative to the pre-industrial level, under a CO_2 doubling scenario and using the mid-range IPCC gray body parameterization? In your opinion, is this is a realistic approach? [HINT: Start by recording temperature (with AGW) for the default values of the solar constant and the Earth's albedo, and double pre-industrial CO_2 using the mid-range gray body IPCC parameterization. This setup produces a 3 °C increase in temperature over the pre-industrial CO_2 levels. Then, increase the Earth's albedo until you find that the temperature (with AGW) decreased by 1°C (ignore *Temperature Change* field, as it does not change). Note that the increase will be very small, so it might be easier to enter values by hand into the box below the slider.]
7. Save your word processing document as either a Microsoft Word or PDF file in the following format:

    PS3_AccessAccountID_LastName.doc (or .pdf).

    For example, student Elvis Aaron Presley's file would be named "PS3_eap1_presley.doc". This naming convention is important, as it will help the instructor match each submission up with the right student!

Submitting your work

- Upload your file to the "Problem Set #3" assignment in Canvas by the due date indicated in the syllabus.

Grading rubric

The instructor will use the general grading rubric for problem sets [116] to grade this activity.

Lesson 4 Summary

In this lesson, we began to explore the use of theoretical models of the climate system. We saw that:

- a simple *zero-dimensional energy balance model* can be used to estimate the surface temperature of the Earth, as well as the response of surface temperatures to changes external (including human-induced perturbations). The model balances the incoming solar radiation

surface temperatures to changes external (including human-induced perturbations). The model balances the incoming solar radiation absorbed at Earth's surface and the outgoing longwave radiation emitted from Earth's surface;
- a simple linear approximation can be used in the zero dimensional EBM to represent the outgoing longwave radiation, leading to a mathematical simplification and a simple formula for global surface temperature;
- using the simplest, *black body* approximation for the outgoing longwave radiation gives a global surface temperature of about 255K, i.e., 18C below freezing—obviously way too cold;
- the *gray body approximation* provides a simple fix to the zero-dimensional EBM that incorporates, at least crudely, the atmospheric greenhouse effect;
- using appropriate values of the *gray body* model coefficients, we can accurately predict both the Earth's surface temperature (roughly 288K, i.e., approximately 15°C), and the response of surface temperatures to perturbations such as increasing greenhouse gas concentrations (roughly 3°C for a doubling of atmospheric CO_2).

This lesson also introduced us to the important concept of equilibrium climate sensitivity—a concept we will encounter again and again throughout this course.

Reminder - Complete all of the lesson tasks!

You have finished Lesson 4. Double-check the list of requirements on the first page of this lesson to make sure you have completed all of the activities listed there before beginning the next lesson.

Lesson 5 - Modeling of the Climate System, part 2

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

About Lesson 5

In this lesson, we will continue with our investigation of climate models. We will investigate more complex models of the climate system than in the previous lesson. We will first investigate a slightly more complex version of the EBM encountered in Lesson 4, where we explicitly insert an atmospheric layer above the Earth's surface. We will consider models that represent the full three-dimensional geometry of the Earth system, and model atmospheric winds and ocean currents, patterns of rainfall, and drought, and other key attributes of the climate system. We will also explore

the concept of 'fingerprint detection'—a method that allows us to compare model predictions against observations to discern whether or not the signal of anthropogenic climate change can already be detected.

What will we learn in Lesson 5?

By the end of Lesson 5, you should be able to:

- describe the hierarchy of theoretical climate models, the underlying assumptions, caveats, and strengths and weaknesses of various climate modeling approaches;
- discuss both the strengths and limitations of current-generation climate models;
- speak to the issue of how climate models have been 'validated';
- assess the relative roles of human vs. natural impacts on climate, based on experiments with climate models;
- assess the state of our current knowledge regarding the equilibrium climate sensitivity of Earth.

What will be due for Lesson 5?

Please refer to the Syllabus for specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 5. Detailed directions and submission instructions are located within this lesson.

- Read:
  - IPCC Fifth Assessment Report, Working Group 1 [2]
  - Summary for Policy Makers [3]
    - D. Understanding the Climate System and its Recent Changes: p. 15-19
  - Dire Predictions, v.2: p. 36-37, 100-101, 110-111, 148-149
- Problem Set #4

Questions?

If you have any questions, please post them to our Questions? discussion forum (not e-mail), located under the Home tab in Canvas. The instructor will check that discussion forum daily to respond. Also, please feel free to post your own responses if you can help with any of the posted questions.

One-Layer Energy Balance Model

We can increase the complexity of the zero-dimensional model by incorporating the atmospheric greenhouse effect in a slightly more realistic manner than is embodied by the *ad hoc* gray body model explored in the previous lecture. We now include an explicit atmospheric layer in the model, which has the ability to absorb and emit infrared radiation.

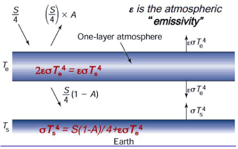


Figure 5.1: One Layer Energy Balance Model.
Credit: M. Mann modification of a figure from Kump, Kasting, Crane "Earth System"

We will approximate the emissivity of Earth's surface as one, that is, we will assume that The Earth's surface emits radiation as a black body. The atmosphere itself has a lower but non-zero emissivity, i.e., it emits a fraction of what a black body would emit at a given temperature. This emissivity is due to property of greenhouse gases within the atmosphere, and we will denote this atmospheric emissivity by ε (not to be confused with the epsilon of previous lessons which was associated with the emissivity of Earth's surface, which we are approximating here as unity!). According to *Kirchhoff's Law*, at thermal equilibrium, the emissivity of a body equals its absorptivity (i.e., the the fraction of incident radiation that is absorbed by the body). Therefore, ε is also a measure of the efficiency of the atmosphere's absorption of any infrared radiation (IR) incident upon it. IR radiation that is not absorbed by the atmosphere is transmitted through it; therefore, $1\text{-}\varepsilon$ is the fraction of incident IR radiation that is transmitted through the atmosphere without being absorbed.

An ε of zero corresponds to no greenhouse effect at all, while an ε of unity corresponds to a perfect IR absorber, i.e., a perfect greenhouse effect. The true greenhouse effect is, of course, somewhere in between, i.e., $0<\varepsilon<1$

We denote the effective *albedo of the Earth system* (i.e., the portion of incoming solar radiation immediately reflected back to space) as *A*, and we will now distinguish between the *atmospheric temperature* T_e (which we will envision as representing the mid-troposphere, somewhere around 5.5 km above the surface where roughly half the atmosphere by mass lies below) and the *surface temperature* T_S.

T_e is related to, but not equivalent to, another quantity known as the *effective radiating temperature*, which we will denote as T_o. T_o is the temperature the Earth would have if it were a black body, i.e., if there were no *greenhouse effect*. It can be thought of as the temperature at the effective height in the atmosphere from which Earth is radiating infrared radiation back to space. In the limit of a perfectly emissive atmosphere ($\varepsilon=1$), as you can verify from our mathematical treatment below, we would have the equality $T_o = T_e$.

You may recall from our earlier discussion (in Lesson 1 [125]) of the vertical structure of the atmosphere, that atmospheric temperatures cool on average roughly $\lambda=6.5^\circ\mathrm{C/km}$ in the troposphere — what is known as the *standard lapse rate*.

For the approximate current value of the solar constant *S* = 1370 W / m^2, we saw in Lesson 4 that the black body temperature, i.e., the effective radiating temperature T_o, is roughly 255 K [126].
Think About It!

*Come up with an answer to this question and then click the words **Reveal answer** below.*

Given that the Earth's average surface temperature is T_s = 288 K, the effective radiating temperature is T_o= 255 K, and the standard lapse rate is $\lambda=6.5^\circ\mathrm{C/km}$, can you determine the effective radiating level in the atmosphere?

Reveal answer.

We can now express the condition of energy balance at each level in our simplified model of the Earth:

1. the top of the atmosphere
2. the atmospheric layer, which we can think of as centered in the mid-troposphere
3. the surface

Balancing incoming and outgoing radiation at the *top of the atmosphere* gives:

$$\frac{S\,(1-A)}{4} = \sigma\varepsilon T_e^4 + (1-\varepsilon)\,\sigma T_S^4$$

(1)

Balancing incoming and outgoing radiation from the *atmospheric layer* gives :

$$\sigma\varepsilon T_S^4 = 2\sigma\varepsilon T_e^4$$

(2)

(Note that short wave radiation is not included in this balance because the atmosphere does not absorb in short wave range.)

Finally, balancing incoming and outgoing radiation at the *surface* gives:

$$\frac{S\,(1-A)}{4} + \sigma\varepsilon T_e^4 = \sigma T_S^4$$

(3)

Solving the system of equations for T_s and T_e gives:

$$T_S^4 = \frac{S\,(1-A)}{[4\sigma\,(1-\varepsilon/2)]}$$

(4)

$$2T_e^4 = T_S^4$$

(5)

Or more simply

$$T_S = \left\{\frac{(1-A)\,S}{[4\sigma\,(1-\varepsilon/2)]}\right\}^{1/4}$$

(6)

$$T_e = \frac{T_S}{2^{1/4}}$$

(7)

Let us use the standard values of *A* = 0.3 and S = 1370 W / m^2.

If we take ε = 0 (which is equivalent to there being no greenhouse effect), we get our original blackbody result T_s = 255 K = -18°C. Too cold!
If we take ε = 1 (which is equivalent to a perfectly IR absorbing atmosphere), we get the result T_s = 303 K = 30°C. Too warm!
However, if we take ε = 0.77 (i.e., the atmosphere absorbs 77% of the IR radiation incident upon it), we get a result, T_s = 288 K = 15°C. Just right!

Using (7) and T_s = 288 K, we also get the result T_e = 242 K. This is modestly lower than the effective radiating temperature T_o = 255 K, indicating that it is found at about 5.5 km — a modestly higher level in the atmosphere than 5.1 km that you calculated earlier.

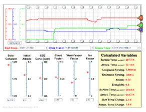
Of course, this model is still rather simplistic. For one thing, it only takes into account short wave and long wave radiation. We haven't accounted for important processes involved in the energy budget of the actual atmosphere and surface, which includes convection, latent heating, and the effect of large-scale motion.

We can nonetheless add *some* further realism to the model by incorporating some of the feedbacks we have discussed [127] previously. In Problem Set 4 you will investigate this slightly more sophisticated [128] version of the standard one-layer model. The model allows for the contribution of

clouds to both the Earth's albedo and the longwave absorptive properties of the atmosphere in a very rough way. It also accounts for the positive ice albedo and water vapor feedbacks in a very rough manner.

Each of the feedbacks in the model will be expressed in the form of a *feedback factor* that you can vary. A *feedback factor* measures the relative magnitude of a feedback in terms of the amplitude of the response relative to the original forcing. If the response is equal in magnitude to the original forcing, there is no feedback, and the feedback factor is zero. If the response is double that of the original forcing, the feedback factor is one. For example, if a warming of 1°C due to CO_2 doubling alone causes an increase in water vapor content that adds an additional equilibrium warming of 2°C, so that the net warming is 3°C, the water vapor feedback factor would be two. Feedback factors can be specified for a particular feedback (e.g., the water vapor feedback), or for the sum over all feedbacks under consideration (e.g., water vapor feedback, ice albedo feedback, and cloud feedback). For example, suppose that the initial 1°C warming, which led to 2°C warming due to water vapor feedback, also led to an increase primarily in low cloud cover, which added a relative cooling of -0.5°C, and a melting of ice, which added an additional relative warming of 1°C. Then the cloud feedback factor would be -0.5, the ice albedo feedback factor would be 1.0, and the net feedback factor would be 2 - 0.5 + 1 = 2.5! Alternatively, we could compute the overall feedback factor by taking the *total warming* (initial 1°C warming + 2°C - 0.5°C + 1°C = 3.5°C) divided by the *initial warming*, *minus one*, i.e., (3.5°C/1°C) - 1 = 2.5. The equilibrium climate sensitivity in this case would be 3.5.

While we have measured the feedback factors i*n terms of the temperature responses*, one could also compute these factors *in terms of the associated radiative forcing*. For example, we know from earlier in the course that the radiative forcing due to CO_2 doubling is roughly 3.7W/m^2. Suppose that the increased greenhouse forcing associated with the water vapor feedback led to an additional downward long wave radiative flux of 7.4 W/m^2 (and let us assume for this example that the other feedbacks are zero). Then the water vapor feedback factor would be 7.4/3.7 which is, again, two. The total downward radiative forcing would be 3.7W/m^2 +7.4W/m^2 =11.1 W/m^2 total downward and the overall feedback factor would be 11.1/3.7 - 1 = 2!



Click on the link to see how our new One-Layer Energy Balance Model [129] works

One-Dimensional Energy Balance Model

There are many ways one can generalize upon the zero-dimensional EBM. As we saw in the previous section, we can try to resolve the additional, *vertical degree of freedom* in the climate system through a very simple idealization—the *one layer* generalization of the zero-dimensional EBM. If for no other reason than the fact that the incoming solar radiation is symmetric with respect to longitude, but varies quite dramatically with latitude, the *latitudinal degree of freedom* is the next most important property to resolve if we wish to obtain further insights into the climate system using a still relatively simple and tractable model.

That brings us to the concept of the *one-dimensional energy balance model*, where we now explicitly divide the Earth up into latitudinal bands, though we treat the Earth as uniform with respect to longitude. By introducing latitude, we can now more realistically represent processes like ice feedbacks which have a strong latitudinal component, since ice tends to be restricted to higher latitude regions.

Recall that we had, for the linearized zero-dimensional gray body EBM, a simple balance [126]:

$C\frac{dT_S}{dt} = \frac{(1-\alpha)S}{4} - A - BT_S$
(1)

where α is the Earth's albedo and A and B are coefficients for the linearized representation of the 4th degree term.

Generalizing the zero-dimensional EBM, we can write a similar radiation and energy balance equation for each latitude band i:

$$C_p\frac{dT_i}{dt} = (1-\alpha_i)\,S_i - A - BT_i$$

(2)

where i represents each latitude band.

We have now introduced some extremely important generalizations. The temperature T_i, albedo α_i , and incoming solar radiation T_S are now functions of latitude, allowing us to represent the disparity in incoming shortwave radiation between equator and pole, and the strong potential latitudinal dependence of albedo with latitude—in particular, when the temperature T_i for a particular latitude zone falls below freezing, we represent the increased accumulation of snow/ice in terms of a higher albedo. The global average temperature T_S is computed by an appropriate averaging of the temperatures for the different latitude bands T_i.

Recall that the disparity in received solar radiation between low and high latitudes leads to lateral heat transport [57] over the surface of the Earth by the atmospheric circulation and ocean currents. In the absence of lateral transport, the poles will become increasingly cold and the equator increasingly warm. Clearly, we must somehow represent this meridional heat transport in the model if we expect realistic results. This can be done through a very crude representation of the process of heat advection through a term that is proportional to the difference between the temperature, $F\,[T_i - T_s]$ where F is some appropriately chosen constant, and T_S is the global average temperature. This term represents processes associated with lateral heat advection that tends to warm regions that are colder than the global average and cool regions that are warmer than the global average.

This gives the final form of our one-dimensional EBM:

$$C_p\frac{dT_i}{dt} + F\,(T_i - T_s) = (1-\alpha_i)\,S_i - A - BT_i$$

(3)

The model is complex enough now that there is no way to simply write down the solution anymore. But we can solve the model mathematically, through a very simple and primitive form of something we will encounter much more of in the future—a *numerical climate model*.

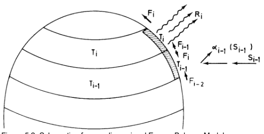


Figure 5.2: Schematic of a one-dimensional Energy Balance Model.
Credit: NYU, Courant Institute of Mathematical Sciences (CIMS) [130]
A Climate Modeling Primer, A. Henderson-Sellers and K. McGuffie, Wiley, pg. 58, (1987)

One of the most important problems that was first studied using this simple one-dimensional model was the problem of how the Earth goes into and comes out of *Ice Ages*. Use the links below to open the demonstration, which is in 3 parts.

- Part One [131]
- Part Two [132]
- Part Three [133]

General Circulation Models

Finally, we come to the so-called *General Circulation Models* or GCMs. GCMs attempt to describe the full three-dimensional geometry of the atmosphere and other components of Earth's climate system. Atmospheric GCMs numerically solve the equations of physics (e.g., dynamics, thermodynamics, radiative transfer, etc.) and chemistry applied to the atmosphere and its constituent components, including the greenhouse gases. In more primitive GCMs (the earlier generation models), the role of the ocean was treated in a very basic way, e.g., as a simple slab of water where only the thermodynamic role of the ocean was accounted for.

Current generation climate models typically include an ocean that plays a far more active role in the climate system. The major current systems are modeled, as is their direct role in transporting heat poleward. When the dynamics of the ocean and its interactions with the atmosphere are explicitly resolved by a climate model, the model is referred to as *Atmosphere-Ocean GCM*, or AOGCM, or sometimes simply a *coupled model*. Most state-of-the-art climate modeling centers today run AOGCMs. In addition, many state-of-the-art climate models today include a detailed description of the hydrological cycle (which couples atmospheric, terrestrial, and ocean reservoirs of water and the flows between these reservoirs) as well as the role of terrestrial biosphere, the continental ice sheets, and even the ocean's carbon cycle and its interactions with the ocean and the atmosphere.

Unlike simpler climate models like EBMs, GCMs and AOGCMS can be used to study a variety of climate attributes other than surface temperature, such as atmospheric temperature profiles, rainfall, atmospheric circulation, ocean circulation, wind patterns, snow and ice distributions, and many other variables that are part of the global climate system.

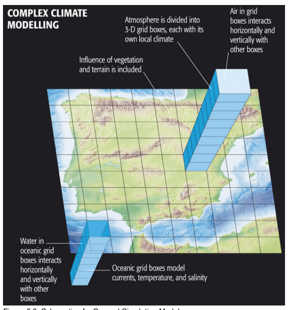


Figure 5.3: Schematic of a General Circulation Model.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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EdGCM

The EdGCM project [134], funded by the U.S. National Science Foundation and spearheaded by scientists associated with NASA Goddard Institute of Space Studies (GISS), uses the GCM originally used in a number of famous experiments (which we will review later in this lesson) by climate scientist James Hansen, Director of GISS.This model was developed in the 1980s and is primitive by modern standards, but it includes much of the important physics that is in current state-of-the-art climate models and it is far less computationally intensive. The scientists at EdGCM have ported the model into a format that can be run on a simple desktop or laptop computer (both PC and Mac). Originally it was free, but to cover expenses for the project, a minor fee is now required for download. Your course author has downloaded EdGCM onto his own laptop (Macbook Pro) and is now going to show you the results of several experiments he has run.

- Part One [135]
- Part Two [136]
- Part Three [137]
- Part Four [138]
- Part Five [139]
- Part Six [140]
- Part Seven [141]
- Part Eight [142]
- Part Nine [143]

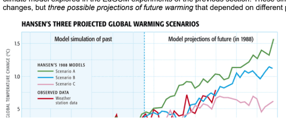
Validating Climate Models

James Hansen
Photo Source: Wikipedia [144]

James Hansen [145] is a well known climate scientist who directs NASA's Goddard Institute for Space Studies. [26]

He was the first climate scientist to testify in the U.S. Congress that human-caused climate change had indeed arrived, back during the hot summer of 1988. Today, as far greater evidence has amassed, his early comments appear especially prescient.

During his 1988 congressional testimony, Hansen showed the results of simulations he had performed using the NASA GISS GCM—the very same climate model explored in the EdGCM experiments of the previous section. These simulations included not only historical simulation of past climate changes, but *three possible projections of future warming* that depended on different possible future fossil fuel utilization scenarios.



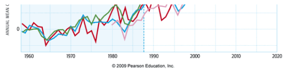


Figure 5.4: Model projections of global temperature by James Hansen in 1988 for three different fossil fuel emissions scenarios compared with the actual temperature observations.
Credit: Mann & Kump, *Dire Predictions, 2008 Dorling Kindersley Limited.*

Yogi Berra is quoted as having once said, "predictions are hard—especially about the future". Indeed, there is no better test than making a prediction about the future, and looking back and seeing how it panned out. This sort of post hoc validation is often done with numerical weather models. It's more difficult to do with climate models, however, because you have to wait not days or weeks, but years to see how the prediction actually measured up.

Hansen's 1988 simulations, in this regard, can be viewed as one of the great validation experiments in climate modeling history. In these experiments, Hansen included a high, medium, and low fossil fuel future emissions scenario, corresponding to the green, blue, and purple curves respectively. As it turns out, our actual fossil fuel emissions scenario during the two decades subsequent to Hansen's 1988 projections, has corresponded most closely to his middle scenario, the blue curve. And as you can see from the subsequent observations (the red curve), his prediction for that scenario quite closely matched the observed warming.

Now, you may have noticed, however, that this model simulation didn't capture the observed multi-year cooling in 1992. Is that a fault of the simulation?

No!—there is no way that James Hansen (or anyone for that matter) could have predicted the eruption of Mt. Pinatubo. And rather than proving a fault with the model, the Pinatubo eruption actually provided Hansen with another key test of the climate models. It takes about 6 months for the volcanic aerosol to spread out around the globe and begin to have a global cooling impact. This gave Hansen about six months to run his model and make a prediction, at the instant Pinatubo erupted. As you can see, he was able to predict quite accurately the short-term cooling of the globe by a bit less than 1°C that would result from this eruption. His model simulation (the black curve below) actually predicted a bit too much cooling (observations shown by the blue curve below). But that, too, wasn't his fault. El Niño events occur randomly in time, and there was no way to know that an extended El Niño event would occur in 1991-1993, offsetting some of the volcanic cooling: As you found in your first problem set, El Niño events warm the globe by about 0.1-0.2°C.

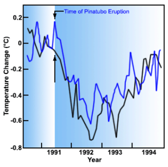


Figure 5.5: Observed cooling following 1991 eruption of Mt Pinatubo vs. cooling predicted in Hansen's climate modeling experiment. [Enlarge with original caption [146]]
Credit: Iowa State Global Change Course [147]

These examples may be the most striking examples of how the models have been validated, but they have been validated in many other more mundane ways. In fact, the various reports of the IPCC include hundreds of pages of 'model validation' (see e.g., chapter 9 on model evaluation [148] in the recent IPCC Fifth Assessment Report) showing the models do a good job capturing the main fluxes of energy and radiative balances, the general circulation of the atmosphere and the major ocean current systems, the amplitude and pattern of the seasonal response to changing patterns of solar insolation, etc.

Detecting Climate Change

So, we have seen in the previous section that climate models have been used to make some very successful predictions in the past, and there is reason to take them seriously. Can we use these models to go a step further than we already have? We have seen in previous lessons that modern-day climate change appears anomalous and without any obvious precedent in the historical past. That alone does not establish that the changes that we are seeing—warming of the Earth's surface, and many other changes—are due to human impacts. Using climate models, we can, however, address this issue of causality. We can use the models to investigate the hypothesis that the observed changes can be explained by *nature alone*, and the alternative hypothesis that they can only be explained by a combination of *human and natural factors*. Investigations employing more than 20 state-of-the-art climate models (see below) show that natural factors alone cannot explain the global temperature record of the past century—including the long-term warming trend—while human factors, combined with the natural factors, can.

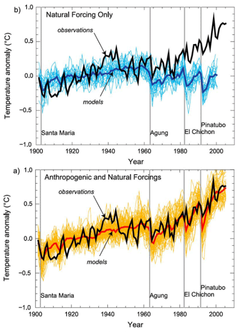


Figure 5.6: Model Simulations of Surface Warming over Past Century Compared with Observations for (top) natural forcing only and (bottom) natural+anthropogenic forcing.
Credit: IPCC, 2007

Some might argue that this alone is not convincing evidence. Perhaps, for example, we simply have the trend in solar output *wrong* and the *true* trend in solar output closely resembles the trend in human impacts (i.e., greenhouse forcing + anthropogenic aerosols). Then we might be misinterpreting the goodness of the fit shown above.

Let us, for argument's sake, accept that criticism. Is there some other type of comparison of observations and model predictions that might be more robust in this situation? Well, we can try to take advantage of the fact that the patterns of response to different forcings might look different. It turns out that the surface expressions are not that different—the surface expression of warming due to solar output increases actually looks a fair amount like the pattern of surface warming due to greenhouse gas increases. The *vertical* patterns of temperature change, however, as we alluded to previously in the course [149], are expected to be quite different. They provide a true *fingerprint* to search for—and indeed, the process of using the expected patterns of response of different forcings to determine which forcings best explain the observed changes is known as *fingerprint detection*.

The vertical pattern of response to increasing greenhouse gas concentrations is one in which the troposphere warms (as we have seen in previous exercises), *but* the stratosphere cools at the expense of this tropospheric warming; greenhouse forcing is a zero sum game and there is no increase in radiation at the top of the atmosphere, but merely a redistribution of energy and radiation within the atmosphere. The vertical pattern of temperature change we would expect for an increase in *solar output*, however, is one in which the *entire atmosphere warms*, from top-to-bottom, as there is an increase in the received radiation at the top of the atmosphere which warms the entire atmospheric column. The pattern of temperature response to an explosive volcanic eruption is yet different from either of these patterns. From comparing the observed patterns of vertical temperature change to model simulations of the responses to each of these different factors, we find that only greenhouse surface warming exhibits the vertical pattern consistent with the model's predicted fingerprint (in fact, there is also impact of ozone depletion on the cooling of the stratosphere, but even after accounting for that effect, the remaining trend can clearly only be accounted for by greenhouse forcing).

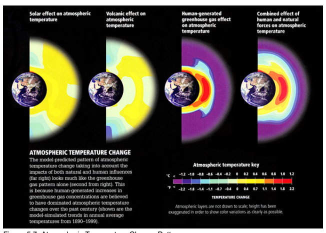


Figure 5.7: Atmospheric Temperature Change Pattern.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Estimating Climate Sensitivity

One of the key unknowns in the behavior of the climate, as we have seen, is the sensitivity—how much warming we can expect in response to a doubling of atmospheric CO_2 concentrations. Current evidence suggests a most likely value of around 3.0°C warming, but there is—as we have seen—a wide range, anywhere from roughly 1.5°C to 4.5°C. Scientists attempt to try to constrain estimates of this key quantity by comparing model simulations with observations.

For example, scientists use models similar to the zero-dimensional EBMs we discussed in Lesson 4, driving them with the estimated changes in both natural factors (volcanoes and solar output) and human factors (greenhouse gas increases and sulphate aerosol emissions). Since the climate sensitivity is simply a parameter that can be changed in the model, scientists can do many simulations using different values of the climate sensitivity, and observe which values yield the best fit with the observations.

Such experiments can be done over the modern period back to the mid 19th century during which observations of global mean temperature are available.

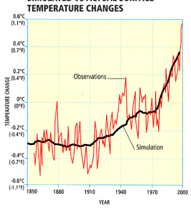

Figure 5.8: Global Temperature vs. Model Simulations During the Modern Instrumental era.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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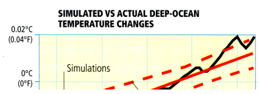
During the shorter period of the past half century when deep ocean temperature observations are available, experiments can be done to compare the model-simulated changes in ocean heat content with those that have been observed.



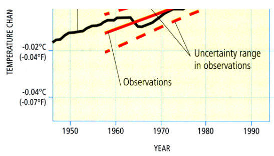


Figure 5.9: Deep Ocean Temperatures vs. Model Simulations During the Past Half Century.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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For the longer period of the past millennium during which temperature changes, as we have seen in Lesson 3 [150], have been documented based on climate *proxy* data—it is possible to compare simulated and observed changes over a longer time period, providing potentially tighter constraints on climate sensitivity. The computer model simulations in this case are driven by longer-term estimates (e.g., from ice core evidence) of natural (volcanic and solar) forcings as well as modern anthropogenic forcing:

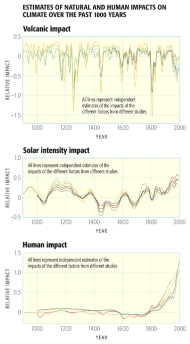

Figure 5.10: Radiative Forcing Estimates Used to Drive Climate Model Simulations of Past Millennium.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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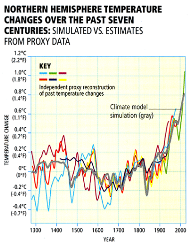

Figure 5.11: Proxy Reconstructions of Northern Hemisphere Temperatures Over Past Millennium Compared Against Model Simulations.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Going further back in time, scientists compare climate model simulations of the cooling during the height of the Last Glacial Maximum (LGM) roughly 21,000 years ago resulting from lowered atmospheric CO_2, increased continental ice cover, and altered patterns of solar insolation, and proxy evidence of ocean surface cooling derived from climate-sensitive surface dwelling organisms trapped in ocean sediment cores.

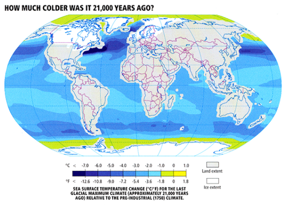

Figure 5.12: Proxy Evidence of Ocean Surface Temperatures During the LGM.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Finally, going even *further* back in time, into the deep geological past, scientists compare model results with geological evidence of past warm and cold periods.

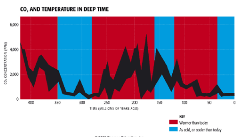

Figure 5.13: Deep Ocean Temperatures vs. Model Simulations During the Past Half Century.
Credit: Mann and Kump, *Dire Predictions: Understanding Global Warming* (DK, 2008)

The overall evidence from all of these different lines of evidence regarding both human-caused and natural climate changes over a broad range of time scales, is that the equilibrium climate sensitivity likely falls within the range of 1.5°C to 4.5°C for CO_2 doubling, with a most likely value of roughly 3°C warming.

Given the full array of available evidence from instrumental and paleoclimate proxy data, and the comparisons of this evidence with theoretical estimates, there is a very low likelihood of either a *trivially small* (e.g., 1.5°C or less) or *extremely high* (greater than 7°C) equilibrium climate sensitivity.

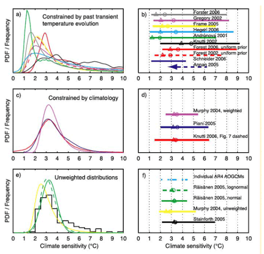

Figure 5.14: Range of Estimates of Equilibrium Climate Sensitivity Based on Studies of (top row) Past Climate Variablity, (middle row) current-day climatological data and (bottom row) Fitting Parameters of Climate Model to Available Climate Observations.
Credit: IPCC 2007

Problem Set #4

Modeling the Earth's Climate Using One-Layer Energy Balance Model

Activity

Note:

For this assignment, you will need to record your work on a word processing document. Your work must be submitted in Word (.doc or .docx) or PDF (.pdf) format so the instructor can open it.

For this activity, you will explore the warming of the surface and the atmosphere due to increases in CO_2 using a one-layer EBM climate model. You will consider/investigate the role of various feedbacks in the climate system (water vapor, ice, and clouds), the influence they have on climate sensitivity, and the impact of the uncertainties in the precise magnitudes of the feedbacks.

Link to The One Layer EBM Application [128]

Directions

1. First, save the Problem Set #4 Worksheet [151] to your computer. You will use this word processing document to electronically record your work in the remaining steps.
   - Save the worksheet to your computer by right-clicking on the link above and selecting "Save link as..."
   - The worksheet is in Microsoft Word format. You can use either Word or Google Docs (free) to work on this assignment. You will submit your worksheet at the end of the activity, so it must be in Word (.doc or .docx) or PDF (.pdf) format so the instructor can open it.
   - Please show your work! When you are explicitly asked to create plots in a question, please cut-and-paste graphics and the output from the screen (e.g., by first printing the output to a pdf file and then directly inserting into the worksheet) to submit along with your discussion and conclusions.
2. Using the online one layer EBM application, double the CO_2 concentrations relative to the pre-industrial level and calculate the climate sensitivity and warming of the mid-troposphere for the following three cases: (A) no feedbacks (i.e., cloud feedback, water vapor feedback, and ice feedback factors all set to zero); (B) mid-range feedback factors (i.e., the default settings of -0.83 for cloud feedback, 2 for water vapor feedback, and 0.5 for ice feedback); (C) high-end estimates of the feedback factors (i.e., the highest settings allowed by the sliders). How does your calculated climate sensitivity range compare with the prevailing range of climate sensitivity estimates? In each case, does the mid-troposphere warm more, the same, or less than the surface? Is this pattern of warming consistent with the predictions by the state-of-the-art climate models? If not, what physics do you think might be missing in our one-layer model?
3. For the doubling of CO_2 for the same three cases (A), (B), and (C) explored in Question 2, please answer the following questions. What are

3. the long wave and short wave forcing, and the total surface forcing (i.e., the sum of the two)? Calculate the overall feedback factor. To do that, first take the ratio of the total surface forcing to the forcing due to the direct radiative impact of CO_2 doubling alone, which simply equals the total forcing in the "no-feedback"□ case (A); then subtract one from this ratio. What are the Earth's albedo and the atmospheric emissivity?

   For cases (B) and (C) is there a change in the Earth's albedo and atmospheric emissivity compared to the "no feedback"□ case (A)? Which feedbacks could be responsible for the observed changes in each case?
4. Save your word processing document as either a Microsoft Word or PDF file in the following format:

   PS4_AccessAccountID_LastName.doc (or .pdf).
   For example, student Elvis Aaron Presley's file would be named "PS4_eap1_presley.doc" This naming convention is important, as it will help me make sure I match each submission up with the right student!

Submitting your work

- Upload your file to the "Problem Set #4" assignment in Canvas by the due date indicated on our Canvas calendar.

Grading rubric

The instructor will use the general grading rubric for problem sets [152] to grade this activity.

Lesson 5 Summary

In this lesson, we further explored the use of theoretical models of the climate system. We found that:

- a generalization of the *zero-dimensional EBM* known as *the one-layer EBM* can be used to provide a more realistic description of the greenhouse effect. This model can be used to estimate both surface temperatures and temperatures of the mid troposphere. It is also possible to study the effect of *feedbacks* using a simple model of this sort;
- a further generalization known as the *one-dimensional EBM* can be used to study the latitudinal dependence of energy balance and temperature distributions. The one-dimensional EBM can be used, among other applications, to try to understand the processes that drive climate into and out of Ice Ages;
- full three-dimensional general circulation models (GCMs) and coupled Atmosphere-Ocean (AOGCM) versions of the the GCM can be used to model the more detailed patterns of climate variability and climate change, and the study not just of temperature changes but other key fields such as precipitation, wind patterns, etc.;
- theoretical climate models have been validated in numerous ways. Predictions of warming made back in the late 1980s have been borne out, and experiments simulating the response to natural events, such as volcanic eruptions, have demonstrated that climate models have the ability to make accurate predictions of the responses of the climate to both natural and human forcings;
- comparisons of model simulations and observations, including so called "fingerprint detection" studies, indicate that natural factors alone cannot explain the observed trends of the past century; only a combination of natural and human factors can explain these trends;
- by comparing model simulations and observations on a variety of timescales, scientists have constrained climate sensitivity—the equilibrium warming expected in response to a doubling of CO_2 concentrations—to lie somewhere within the range of 1.5 to 4.5°C, with a most likely estimate of around 3°C warming.

Reminder - Complete all of the lesson tasks!

You have finished Lesson 5. Double-check the list of requirements on the first page of this lesson [153] to make sure you have completed all of the activities listed there before beginning the next lesson.

Lesson 6 - Carbon Emission Scenarios

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

About Lesson 6

Now that we have explored the underlying workings of the climate system, experimented with actual climate models and validated their predictions, we are in a position to use climate models to make projections of future climate change. Before we can project human-caused climate changes, however, we must consider the various plausible scenarios for future human behavior, and resulting greenhouse gas emissions pathways.

What will we learn in Lesson 6?

By the end of Lesson 6, you should be able to:

- discuss the range of hypothetical pathways of future greenhouse gas emissions;
- distinguish between the concepts of CO_2 and CO_2 *equivalent* emissions;
- explain the *Kaya Identity;*
- explain the concept of *stabilization* of greenhouse gas concentrations; and
- discuss the *wedges* concept for controlling greenhouse gas emissions.

What will be due for Lesson 6?

Please refer to the Syllabus for specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 6. Detailed directions and submission instructions are located within this lesson.

- Read
  - IPCC Fifth Assessment Report, Working Group 1 [2]
  - Summary for Policy Makers [3]□
    - E Future Global and Regional Climate Change: p. 19-20
    - E.8 Climate Stabilization, Climate Change Commitment and Irreversibility: p. 27-29
    - Box SPM.1 Representative Concentration Pathways: p. 29
  - Dire Predictions, v.2: p. 92-93

- Begin Project #1: Design your own fossil fuel emissions scenario that would limit future warming by 2100 to 2.5°C relative to pre-industrial levels.
- Participate in Lesson 6 discussion: Carbon Emission Scenarios

Questions?

If you have any questions, please post them to our *Questions?* discussion forum (not e-mail), located under the Home tab in Canvas. The instructor will check that discussion forum daily to respond. Also, please feel free to post your own responses if you can help with any of the posted questions.

'SRES' Scenarios and 'RCP' Pathways

Scientists attempt to create scenarios of future human activity that represent plausible future greenhouse emissions pathways. Ideally, these scenarios span the range of possible future emissions pathways, so that they can be used as a basis for exploring a realistic set of future projections of climate change.

In previous IPCC assessments, the most widely used and referred-to family of emissions scenarios were the so-called SRES scenarios (for Special Report on Emissions Scenarios) that helped form the basis for the IPCC Fourth Assessment Report. These scenarios made varying assumptions ('storylines') regarding future global population growth, technological development, globalization, and societal values. One (the A1 'one global family' storyline chosen by Michael Mann and Lee Kump in version 1 of *Dire Predictions*) assumed a future of globalization and rapid economic and technological growth, including fossil fuel intensive (A1FI), non-fossil fuel intensive (A1T), and balanced (A1B) versions. Another (A2, 'a divided world') assumed a greater emphasis on national identities. The B1 and B2 scenarios assumed more sustainable practices ('utopia'), with more global-focus and regional-focus, respectively.

Let us now directly compare the various SRES scenarios both in terms of their annual rates of carbon emissions, measured in gigatons (Gt) of carbon (1Gt = 10^{12} tons), and the resulting trajectories of atmospheric CO_2 concentrations. Getting the concentrations actually requires an intermediate step involving the use of simple model of ocean carbon uptake, to account for the effect of oceanic absorption of atmospheric CO_2.

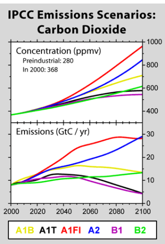

Figure 6.1: Estimated CO_2 concentrations (top) and Annual Carbon Emissions (bottom) for the Various IPCC SRES Scenarios.
Credit: Robert A. Rohde / Global Warming Art [154]

We can see from the above comparison how various trajectories of our future carbon emissions translate to atmospheric CO_2 concentration trajectories. From the point of view of controlling future CO_2 concentrations, these graphics can be quite daunting. Depending on the path chosen by society, we could plausibly approach CO_2 concentrations that are quadruple pre-industrial levels by 2100. Even in the best case of the SRES scenarios, B1, we will likely reach twice pre-industrial levels (i.e., around 550 ppm) by 2100. And to keep CO_2 concentrations below this level, we can see that we have to bring emissions to a peak by 2040, and ramp them down to less than half current levels by 2100.

You might wonder, what scenario do we actually appear to be following? Over the past ten years, observed emissions have actually been close to the most carbon intensive of the SRES scenarios—A1FI. This gives you an idea of how challenging the problem of stabilizing carbon emissions at levels lower than twice pre-industrial actually is.

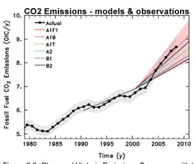

Figure 6.2: Observed Historic Emissions Compares with the Various IPCC SRES Scenarios.
Credit: The Copenhagen Diagnosis [155]

One problem with the SRES scenarios—indeed, a fair criticism of them—is that they do not explicitly incorporate carbon emissions controls. While some of the scenarios involve storylines that embrace generic notions of sustainability and environmental protection, the scenarios do not envision explicit attempts to stabilize CO_2 concentrations at any particular level. For the Fifth Assessment Report, a new set of scenarios, called Representative Concentration Pathways (RCPs), was developed. They are referred to as pathways to emphasize that they are not definitive, but are instead internally consistent time-dependent forcing projections that could potentially be realized with multiple socioeconomic scenarios. In particular, they can take into account climate change mitigation policies to limit emissions. The scenarios are named after the approximate radiative forcing relative to the pre-industrial period achieved either in the year 2100, or at stabilization after 2100. They were created with 'integrated assessment models' that include climate, economic, land use, demographic, and energy-usage effects, whose greenhouse gas concentrations were then converted to an emissions trajectory using carbon cycle models.

The RCP2.6 scenario peaks at 3.0 W / m^2 before declining to 2.6 W / m^2 in 2100, and requires strong mitigation of greenhouse gas concentrations in the 21st century. The RCP4.5 and RCP6.0 scenarios stabilize after 2100 at 4.2 W / m^2 and 6.0 W / m^2, respectively. The RCP4.5 and SRES B1 scenarios are comparable; RCP6.0 lies between the SRES B1 and A1B scenarios. The RCP8.5 scenario is the closest to a 'business as usual' scenario of fossil fuel use, and has comparable forcing to SRES A2 by 2100.

In all RCPs global population levels off or starts to decline by 2100, with a peak value of 12 billion in RCP8.5. Gross domestic product (GDP) increases in all cases; of note, the RCP2.6 pathway has the highest GDP, though it has the least dependence on fossil fuel sources. Carbon dioxide emissions for all RCPs except the RCP8.5 scenario peak by 2100.

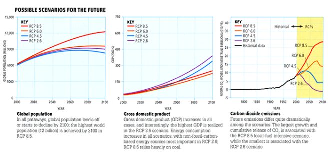

Figure 6.3a: RCP Global Population Scenarios
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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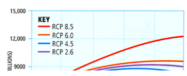


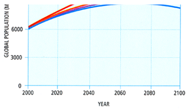


Figure 6.3b: RCP Global Population Scenarios
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*

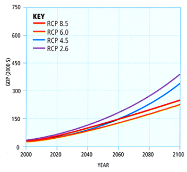


Figure 6.4: RCP Gross Domestic Product Scenarios.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*

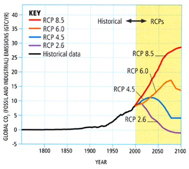


Figure 6.5: RCP Carbon Dioxide Emission Scenarios.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*

As with the SRES scenarios, however, stabilizing CO_2 concentrations requires not just preventing the increase of emissions, but reducing emissions. This leads naturally to our next topic—the topic of *stabilization scenarios*.

Stabilizing CO2 Concentrations

Before we proceed, it is useful to cover a few more important details. You may recall from an earlier lesson [156] that the radiative forcing due to a given increase in atmospheric CO_2 concentration, ΔF_{CO_2} , can be approximated as:

$$\Delta F_{CO_2} = 5.35 \ln\left(\frac{[CO_2]}{[CO_2]_0}\right)$$

where $[CO_2]_0$ is the initial concentration and $[CO_2]$ is the final concentration. This gives a forcing for doubling of CO_2 from pre-industrial values (i.e., $[CO_2]_0$ = 280 ppm and $[CO_2]$ = 560 ppm) of just under $4\ Wm^{-2}$. Given the typical estimate of climate sensitivity we discussed during the past two lessons, we know that this forcing translates to about 3°C warming. That means, we get about 0.75°C warming for each Wm^{-2} of radiative forcing.

Think About It!

*Come up with an answer to this question and then click the words **Reveal answer** below.*

Thus far, CO_2 has increased from pre-industrial levels of 280 ppm to current levels of around 400 ppm. Based on the relationships above, what radiative forcing and global mean temperature increase would you expect in response to our behavior so far?

Reveal answer.

If you successfully answered the question above, you know that the CO_2 increases so far should have given rise to 1.4°C warming of the globe. Yet we have only seen about 0.8°C warming. Are the theoretical formulas wrong? Did we make a mistake? Actually, it is neither. First of all, we know that it takes decades for the climate system to equilibrate to a rise in atmospheric CO_2, so we have not yet realized the expected *equilibrium warming* indicated by the *equilibrium climate sensitivity.* Models indicate that there is as much as another 0.5°C of warming still in the pipeline, due to the CO_2 increases that have taken place already. That alone would essentially explain the 0.6°C discrepancy between the warming we expect, and the lesser warming we've observed.

However, we have forgotten two other things that—as it happens—roughly cancel out! First of all, CO_2 is not the only greenhouse gas whose concentrations we have been increasing through industrial and other human activities. There are other greenhouse gases—methane, nitrous oxide, and others—whose concentrations we have increased, and whose concentrations are projected to continue to rise in the various SRES and RCP scenarios we have examined.

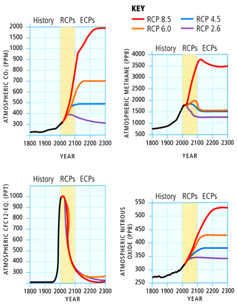


Figure 6.6: Greenhouse Gas Levels Resulting from Various Emissions Scenarios.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*

We need to account for the effect of all of these other greenhouse gases. We can do this using the concept of CO_2 equivalent (CO_2_eq). CO_2_eq is the concentration of CO_2 that would be equivalent, in terms of the total radiative forcing, to a combination of all the other greenhouse gases. If we take into account the rises in methane and other anthropogenic greenhouse gases, then the net radiative forcing is equivalent to having increased CO_2 to a substantially higher, roughly 485 ppm! In other words, the current value of CO_2_eq is 485 ppm. This fact has caused quite a bit of confusion, leading some commentators (see this RealClimate article [157]) to incorrectly sound the alarm that it is already too late to stabilize CO_2 concentrations at 450 ppm and, hence, to avoid breaching the targets that have been set by some as constituting dangerous anthropogenic interference with the climate (see this article by Michael Mann [122] for a discussion of these considerations).

Nonetheless, if CO_2_eq has reached 485ppm, does that mean that we are committed to the net warming that can be expected from a concentration of 485 ppm CO_2? Well, yes and no. The other thing we have left out is that greenhouse gases are not the only significant anthropogenic impact on the climate. We know that the production of sulphate and other aerosols has played an important role, cooling substantial regions of the Northern Hemisphere continents, in particular, during the past century. The best estimate of the impact of this anthropogenc forcing, while quite uncertain, is roughly -0.8 W/m^2 of forcing, which is equivalent—in this context—to the contribution of *negative* 60 ppm of CO_2. If we add -60 ppm to 485 ppm we get 425 ppm—which is closer to the current actual CO_2 concentration of 400 ppm. So, in other words, if we take into account not only the effect of all other greenhouse gases, but also the offsetting cooling effect of anthropogenic aerosols, we end up roughly where we started off, considering only the effect of increasing atmospheric CO_2 concentration through fossil fuel burning.

It is, therefore, a useful simplification to simply look at atmospheric CO_2 alone as a proxy for the total anthropogenic forcing of the climate, but there are some important caveats to keep in mind:

> (1) the various scenarios assume that the sulphate aerosol burden remains unchanged. If we instead choose to clean up the atmosphere to the point of scrubbing all current sulphate aerosols from industrial emissions, we are left with the *faustian bargain* [158] of experiencing the additional climate change impacts of a sudden effective increase of atmospheric CO_2 of 60 ppm;
> (2) not all greenhouse gases are created the same—some, such as methane, have far shorter residence times in the atmosphere (timescale of years) than does CO_2, which persists for centuries.

That means that there is a far greater future climate change commitment embodied in a scenario of pure CO_2 emissions than the same CO_2 equivalent emissions consisting largely of methane. This has implications for the abatement strategies we will discuss later in the course.

These limitations notwithstanding, let us now consider the impact of various pure CO_2 scenarios. Let us focus specifically on scenarios that will stabilize atmospheric CO_2 at some particular level, i.e., so-called *stabilization scenarios*. Invariably, these scenarios involve bringing annual emissions to a peak at some point during the 21st century, and decreasing them subsequently. Obviously, the higher we allow the concentrations to increase and the later the peak, the higher the ultimate CO_2 concentration is going to be. The various possible such scenarios are shown below in increments of 50 ppm. If we are to stabilize CO_2 concentrations at 550 ppm, we can see that CO_2 emissions should be brought to a peak of no more than 8.7 gigatons of carbon per year, by around 2050, and reduced below 1990 levels (i.e., 6 gigatons carbon per year) by 2100. For comparison, as we saw earlier that current emissions are at roughly 8.5 gigatons per year and rising at the rate of the carbon-intensive A1FI SRES emissions scenario, so we are already "behind the curve" so to speak, even for 550 ppm stabilization.

For 450 ppm stabilization, the challenge is far greater. According to the figure below, we would have had to bring emissions to a peak before 2010 at roughly 7.5 gigatons per year, and lower them to roughly 4 gigatons per year (i.e., 33% below 1990 levels) by 2050. Obviously, that train has already left the station. Alternatively, the RCP2.6 pathway is an example of a 450 ppm stabilization scenario consistent with where we are now, that involves bringing emissions to a peak within the next decade below 10 gigatons per year, and reducing them far more dramatically, to near zero 80% by 2100 through various mitigation policies. With every year we continue with *business-as-usual* carbon emissions, achieving a 450 ppm stabilization target becomes that much more difficult, and involves far greater reduction of emissions in future decades. It is for this reason that the problem of greenhouse gas stabilization has been referred to by some scientists as a problem with a very large procrastination penalty [159].

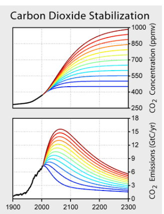


Figure 6.7: Annual CO_2 emissions and Resulting CO_2 concentrations for Various Stablization Scenarios.
Credit: Robert A. Rohde / Global Warming Art [154]

The "Kaya Identity"

We can actually play around with greenhouse gas emissions scenarios ourselves. To do so, we will take advantage of something known as the *Kaya Identify* [160]. Technically, the identity is just a definition, relating the quantity of annual carbon emissions to a factor of terms that reflect (1) *population*, (2) relative (i.e., per capita) *economic production*, measured by annual GDP in dollars/person, (3) *energy intensity,* measured in terawatts of energy consumed per dollar added to GDP, and (4) *carbon efficiency*, measured in gigatons of carbon emitted per terawatt of energy used. Multiply these out, and you get gigatons of carbon emitted. If the other quantities are expressed as a percentage change per year, then the carbon emissions, too, are expressed as a percentage change per year, which, in turn, defines a future trajectory of carbon emissions and CO_2 concentrations.

Mathematically, the Kaya identity is expressed in the form:

$$F = P * (G/P) * (E/G) * (F/E)$$

where

> F is global CO_2 emissions from human sources
> P is global population
> G is world GDP
> E is global energy consumption

By projecting the future changes in *population (P)*, *economic production* (G/P) , *energy intensity* (E/G) , and *carbon efficiency* (F/E) , it is possible to make an informed projection of future *carbon emissions* (F) . Obviously, *population* is important as, in the absence of anything else, more people means more energy use. Moreover, *economic production* measured by GDP per capita plays an important role, as a bigger economy means greater use of energy. The *energy intensity* term is where technology comes in. As we develop new energy technologies or improve the efficiency of existing energy technology, we expect that it will take less energy to increase our GDP by and additional dollar, i.e., we should see a decline in energy intensity. Last, but certainly not least, is the *carbon efficiency*. As we develop and increasingly switch over to renewable energy sources and non-fossil fuel based energy alternatives and improve the carbon efficiency of existing fossil fuel sources (e.g., by finding a way to extract and sequester CO_2), we can expect a decline in this quantity as well, i.e., less carbon emitted per unit of energy production.

Fortunately, we do not have to start from scratch. There is a convenient online calculator [161] here, provided courtesy of David Archer of the University of Chicago (and a RealClimate [162] blogger). Below a brief demonstration of how the tool can be used. After you watch the demonstration, use the link provided above to play around with the calculator yourself.

- Part One [163]
- Part Two [164]
- Part Three [165]
- Part Four [166]

The "Wedges" Concept

An increasingly widespread approach to characterizing greenhouse gas emissions reductions is the so-called *Wedges concept* [167] introduced by Princeton researchers a few years ago. The concept is relatively straightforward. First, one defines the current path of *business-as-usual* emissions. We can think of that ramp-like path as defining a *stabilization triangle*, as shown below.

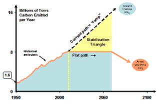


Figure 6.8: Schematic of "Wedges" Approach to Defining Carbon Emissions Reduction Strategies.
Credit: Carbon Mitigation Initiative, Princeton University [168]

Based on the past one to two decades, the business-as-usual pathway corresponds to an increase of about 1.5 gigatons per decade [169]—which, if we extrapolate linearly, amounts to about 7 gigatons of carbon emissions over the next 50 years. The stabilization triangle can thus be split into 7 "wedges" that each represent 1 gigaton of carbon over the next 50 years. The first step to stabilizing greenhouse gas concentrations is to freeze annual emissions so that they do not rise any further. To accomplish this, we would need to replace 7 gigaton wedges of projected greenhouse gas emissions that would be required to meet the forecasted business-as-usual global energy requirements over the next 50 years. The individual wedges could be derived from greater energy efficiency, decreased reliance on fossil fuels, new technologies aimed at sequestration of CO_2, etc.

Of course, as we have seen from our discussion of stabilization scenarios [170], simply freezing greenhouse emissions at current levels is not adequate to stabilize concentrations. The emissions must be decreased, eventually to zero, or at least close enough to zero so that they are balanced by the natural rates of uptake of carbon from the ocean and biosphere. So, the wedge approach must be supplemented by an actual decrease in emission rates. In one idealization of the approach, the wedges are used to freeze greenhouse annual emissions for 50 years, after which technological innovations that have been developed over the intervening half century presumably make the problem of fully phasing out fossil fuel-based energy more tractable, and emissions can be reduced over the subsequent 50 years as necessary to avoid breaching, e.g., twice pre-industrial CO_2 levels. Alternatively, more additional wedges, beyond the original 7, can be used, to not only freeze annual CO_2 emissions at current levels during the next 50 years, but instead, bring them down.

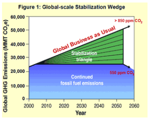


Figure 6.9: Diagram Showing the Individual "Wedges" in the Wedge Strategy.
Credit: EPA [171]

The wedge concept can be generalized beyond the global CO_2 stabilization problem. For example, the U.S. EPA has introduced *wedge-based* plan for reducing emissions in the U.S. transportation sector as a means of mitigating this important current contribution to U.S. greenhouse gas emissions.

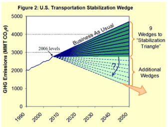


Figure 6.10: Application of Wedge Concept to Greenhouse Emission Reductions in the U.S. Transportation Sector.
Credit: EPA [172]

The Wedge Concept is an increasingly popular way to go about achieving the required greenhouse gas emissions in the decades ahead, by thinking about each of the individual mitigation approaches that might buy us a wedge, or some fraction of a wedge, of reductions. It is a way to think about how to take a seemingly intractable problem and break it up into many smaller, potentially tractable problems which collectively can help civilization achieve the daunting emissions reductions necessary for avoiding potentially dangerous climate change.

Project #1: Fossil Fuel Emissions

Scenario for Limiting Future Warming

Climate Change mitigation is an example of the need for *decision making in the face of uncertainty*. We must take steps today to stabilize greenhouse gas concentrations if we are to prevent future warming of the globe, despite the fact that we do not know precisely how much warming to expect. Furthermore, it is a problem of *risk management*. We do not know precisely what potential impacts loom in our future, and where the threshold for dangerous anthropogenic impacts on the climate lies. Just like in nearly all walks of life, we must make choices in the face of uncertainty, and we must decide precisely how risk averse we are. Most homeowners have fire insurance, yet they don't expect their homes to burn down. They simply want to hedge against the catastrophe if it does happen. We can, in an analogous manner, think of climate change mitigation as

hedging against dangerous potential impacts down the road. This project aims to integrate a number of themes we have already explored—energy balance and climate modeling, and our current lesson on carbon emissions scenarios—to quantify how to go about answering critical questions like, "How do we go about setting emissions limits that will allow us to hedge against the possibility of dangerous anthropogenic impacts (DAI) on our climate?"

Activity

Note:

For this assignment, you will need to record your work on a word processing document. Your work must be submitted in Word (.doc or .docx), or PDF (.pdf) format so the instructor can open it.

For this project, you will design your own fossil fuel emissions scenario that would **limit future warming by the year 2100 to 2.0°C relative to the pre-industrial level**.

Directions

1. Defining the threshold for DAI with the climate is a value judgment, as much as a scientific one. The European Union has defined 2°C warming relative to pre-industrial conditions to be the threshold of DAI (and this has been adopted in the recent Paris agreement). Use the zero-dimensional EBM [118] with the standard, i.e., *mid-range* values of the *gray body parameters* (and default settings for solar constant and albedo) to estimate the CO_2 concentration for which we would achieve such warming when equilibrium is reached. [*Hint*: you already did this particular calculation in Problem Set #3]
2. As discussed above, where we choose to set our emissions limits is also a matter of risk management. How risk averse are we? Let us assume that we want to limit the chance of exceeding DAI to only 5%, which is a typical threshold used in risk abatement strategies. Let us also assume that the *low end* and *high end* IPCC sensitivities define our best estimate of the 90% confidence interval for climate sensitivity, that is, there is a 5% chance that the true sensitivity is lower than the *low end* value, and a 5% chance that the true sensitivity is higher than the *high end* value. Now, revisit the calculation that you did in Step 1, and use *high-end* setting of the *gray body parameters* to calculate the highest CO_2 concentration that we can reach and still keep the chance of exceeding DAI below 5%.
3. That was the easy part! Now is where things really get interesting. You have to come up with *your own emissions scenario* to stabilize CO_2 concentrations below the dangerous CO_2 threshold that you calculated in Step 1 (i.e., using the mid-range values). Recall that based on Step 2 we are not being particularly risk-averse under this assumption. By stabilizing, we will mean that by the year 2100 the CO_2 concentration curve should be flat or nearly flat, indicating that any peak in CO_2 concentrations has been reached before 2100 - and that the value of that peak should not exceed the threshold found in Step 1. You should make use of the on-line Kaya calculator [161] that we examined earlier in this lesson. Begin with the default settings in the Kaya calculator to figure out what baseline emissions scenario you are starting with and how much you need to do to achieve the required reduction in emissions to stabilize the concentration (given by the 'pCO2' curve when you choose 'ISAM pCO2' in the selector for the display on the right). You then have several knobs you can tune to achieve an alternative emissions scenario; specifically, you can do the following.
   - (1) You can take measures to control global *population growth*, which are consistent with the spread of the population trajectories [173] of the various SRES scenarios (e.g., A1, A2, B1, B2); though note that you cannot set the limit below the current global population of 7.4 billion.
   - (2) You can change the *rate of economic expansion* (measured by GDP per capita) by up to 75% from the default value of 1.6% per year; that is, you can choose a value within the range of 0.4 to 2.8 % per year.
   - (3) You may change the rate of decline in *energy intensity* (measured in Watts per dollar) through new technology and/or the improvement of existing technology. You are permitted to change this value by 100% form its default setting of -1% per year, that is you can choose a value within the range of 0 to -2 % per year.
   - (4) You may change the decline in *carbon intensity* by 100% from the default value of -0.3 % per year, that is you can choose a value within the range of 0 to -0.6 % per year. This would would reflect efforts to shift from the current reliance on fossil fuel sources, or the improved carbon efficiency of fossil fuel energy sources, e.g., through sequestration of CO_2, etc.

   Through some combination of tuning these knobs within the indicated ranges, it should be possible to stabilize CO_2 concentrations at the necessary threshold level. (The concentration time-evolution curves corresponding to different specific stabilization scenarios are shown in the 'ISAM pCO2' plot.)
4. As you write up your results, please discuss the reasoning that you used in arriving at the various choices for the factors in the Kaya identity, i.e., provide justification for why your choices reflect plausible policies that governments could in principle implement to achieve the necessary reductions. If your projections depart from what might be expected based on an extrapolation of past historic trends, provide some justification. Your discussion here might, for example, be guided by the plot of the required amounts of carbon-free energy to meet the requirements of your scenario, which is provided by the Kaya calculator tool. You will probably want to do some additional background reading. A good place to start would be the original 2001 IPCC report on SRES scenarios: Special Report on Emissions Scenarios [174].
5. Save your word processing document as either a Microsoft Word or PDF file in the following format:

   Project1_AccessAccountID_LastName.doc (or .pdf).
   For example, student Elvis Aaron Presley's file would be named "P1_eap1_presley.doc"—This naming convention is important, as it will help the instructor match each submission up with the right student.

Submitting your work

- Upload your file to the Project 1 assignment in Canvas by the due date indicated in the Syllabus.

Grading rubric

The instructor will use the general grading rubric for problem sets [152] to grade this project.

Lesson 6 Discussion

Activity

Directions

Please participate in an online discussion of the material presented in Lesson 6: Carbon Emission Scenarios.

This discussion will take place in a threaded discussion forum in Canvas (see the Canvas Guides [18] for the specific information on how to use this tool) over approximately a week-long period of time. Since the class participants will be posting to the discussion forum at various points in time during the week, you will need to check the forum frequently in order to fully participate. You can also subscribe to the discussion and receive e-mail alerts each time there is a new post.

Please realize that a discussion is a group effort and make sure to participate early in order to give your classmates enough time to respond to your posts.

Post your comments addressing some aspect of the material that is of interest to you and respond to other postings by asking for clarification, asking a follow-up question, expanding on what has already been said, etc. For each new topic you are posting, please try to start a new discussion thread with a descriptive title, in order to make the conversation easier to follow.

Suggested Topics

- Discuss the concept of SRES scenarios and RCP pathways. How are these scenarios used for projecting future climate change? Given what we know about the current greenhouse emissions, which scenarios and/or pathways appear to best represent the real world?
- Discuss potential reasons for the switch from SRES scenarios to RCP pathways in the latest IPCC report. How useful do you think the change was?
- What is the difference between CO_2 emissions and CO_2 equivalent emissions?
- Atmospheric CO_2 increases observed so far should have given rise to 1.3°C warming, but we have only seen about 0.8°C warming. Why?
- What are stabilization scenarios?
- Do you find the Wedges Concept a useful tool for characterizing greenhouse gas emissions reductions?

Submitting your work

1. Go to Canvas.
2. Go to the Home tab.
3. Click on the *Lesson 6 discussion: Carbon Emission Scenarios.*
4. Post your comments and responses.

Grading criteria

You will be graded on the quality of your participation. See the online discussion grading rubric [175] for the specifics on how this assignment will be graded. Please note that you will not receive a passing grade on this assignment if you wait until the last day of the discussion to make your first post.

Lesson 6 Summary

In this lesson, we looked at the science underlying greenhouse gas emissions scenarios. We learned that:

- up through the Fourth Assessment Report, the IPCC employed, for the purpose of projecting future greenhouse gas concentrations, a set of emissions scenarios, known as the SRES scenarios. These scenarios reflect a broad range of alternative assumptions about how future technology, economic growth, demographics, and energy policies will evolve over the next century, and, therefore, plausibly reflect the diversity of potential future global greenhouse emissions pathways;
- the SRES scenarios embody a range of projected increases in atmospheric CO_2 by 2100 from a lower end of approximately doubling the

pre-industrial levels to reach 550 ppm (B1) to a near quadrupling of pre-industrial levels (A1FI). Current emissions place us on a pathway close to the upper-end A1FI scenario;
- in the Fifth Assessment Report, the IPCC switched to the use of Representative Concentration Pathways, or RCPs. These pathways (RCP 2.6, RCP 4.5, RCP 6.0 and RCP 8.5) were chosen to be representative scenarios named for their total radiative forcing in the year 2100 (in watts per meter squared), and reflect a range of policies, from strong mitigation (RCP 2.6) to approximately business-as-usual (RCP 8.5);
- the s*tabilization scenarios* are designed to stabilize atmospheric CO_2 concentrations at a particular level. The lower the desired stabilization level, the lower and sooner the peak in emissions must be. To stabilize below twice the pre-industrial levels, emissions must be brought to a peak within the next few decades and rapidly brought down by the end of the century, falling below 1990 levels by mid-century. To stabilize below 450 ppm, CO_2 levels must be brought to a peak within the next decade, and brought down to 80% below 1990 levels by mid-century;
- an increasingly widely used approach to defining the required carbon emissions reductions is the *Wedge* approach. This approach involves freezing emissions at current rates by offsetting projected business-as-usual emissions over the next 50 years (roughly 7 gigatons), envisioned, e.g., as 7 strategies for 1 gigaton carbon emission reductions. After 50 years, emission rates are brought down, but how abruptly and rapidly depends on the stabilization targets desired. Additional wedges can be used to achieve lower stabilization targets by bringing down, rather than freezing, annual carbon emission rates over the next 50 years.

Reminder - Complete all of the lesson tasks!

You have finished Lesson 6. Double-check the list of requirements on the first page of this lesson [176] to make sure you have completed all of the activities listed there before beginning the next lesson.

Lesson 7 - Projected Climate Changes, part 1

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

About Lesson 7

With plausible greenhouse gas emissions scenarios now in hand, we are ready to begin looking at future climate change projections. We will start out by looking at the basic attributes of the projections—changes in surface temperature, atmospheric and oceanic circulation changes, patterns of rainfall and drought, and the climate mechanisms that may influence climate changes at regional spatial scales.

What will we learn in Lesson 7?

By the end of Lesson 7, you should be able to:

- assess the impact of hypothetical pathways of future greenhouse gas emissions on global temperatures in the context of the estimated uncertainties; and
- assess potential impacts of projected climate changes on patterns of rainfall and drought patterns, ocean and atmospheric circulation, and modes of climate variability.

What will be due for Lesson 7?

Please refer to the Syllabus for specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 7. Detailed directions and submission instructions are located within this lesson.

- Take Quiz #2.
- Read:
    - IPCC Fifth Assessment Report, Working Group 1 [2]
    - Summary for Policy Makers [3], Future Global and Regional Climate Change
        - E.1 Atmosphere: Temperature: p. 20
        - E.2 Atmosphere: Water Cycle: p. 20-23
        - E.3 Atmosphere: Air Quality: p. 24
    - Dire Predictions, v.2: p. 98-103

Questions?

If you have any questions, please post them to our *Questions?* discussion forum (not e-mail) located under Home tab in Canvas. The instructor will check that discussion forum daily to respond. Also, please feel free to post your own responses if you can help with any of the posted questions.

Surface Temperature Changes

When it comes to climate change projections, the most obvious first thing to look at is the increases in global mean temperature projected by the climate models. When we do that, we are immediately confronted with two major uncertainties each of a fundamentally different nature.

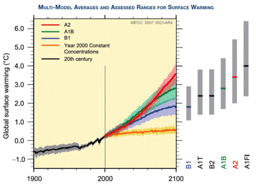


Figure 7.1: Model Projections of Future Warming Under Various Emissions Scenarios.
Credit: IPCC, 2007

The first of the two uncertainties is the *scenario uncertainty*, and is represented by the different families of color-shaded regions in the graph below. It corresponds to the uncertainty in what pathway of future behavior *we* will follow. That uncertainty, in a crude sense, is spanned by the various SRES scenarios explored in the previous lesson. In reality, this set of scenarios alone implies greater constraint on the true spread of potential future pathways of anthropogenic activity, since there are numerous wild-cards, including (i) future anthropogenic aerosol emissions and (ii) potential carbon cycle feedbacks which may accelerate the rate at which the airborne fraction of CO_2 increases with future emissions. That having been said, it is likely that a lower bound corresponding to fixed CO_2 concentrations and an upper bound specified by the A1FI scenario, reasonably brackets the range of future emissions pathways that human civilization will choose to follow.

The second of the two uncertainties is the *physical uncertainty*, and it corresponds to the width of each of the shaded regions (the width of the shading indicates the one standard error range among the 20+ models used in the IPCC assessment; the wider gray bars shown on the right indicate the full range of warming over all 20+ models). Much of this uncertainty comes from the previously discussed uncertainty in cloud radiative feedbacks. On average, as we know from our previous lesson, cloud radiative feedbacks are estimated to be negative. The uncertainty, however, is huge. Among the 20+ models used in the IPCC assessment, the cloud radiative feedback for CO_2 doubling varies anywhere from around $-2W/m^2$ (offsetting roughly half of the direct radiative forcing by the CO_2 increase) to nearly $+2W/m^2$ (adding nearly half of the radiative forcing due to the CO_2 increase alone).

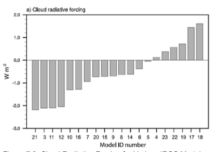


Figure 7.2: Cloud Radiative Forcing for Various IPCC Models.
Credit: IPCC, 2007

Collectively, the various scenarios and their physical uncertainty ranges span a very large spread of projected warming for the next century. In the most optimistic of scenarios—indeed, an arguably unrealistic scenario—where we could manage to keep CO_2 fixed at the year 2000 concentration (this would require immediate cession of all activities—including fossil fuel burning, deforestation, etc.—contributing to anthropogenic CO_2 emissions), warming nonetheless persists for decades owing to the "commitment to warming" we investigated in the previous lesson in our EdGCM experiments. This is warming already in the pipeline but not yet realized because of the delayed response of ocean warming to greenhouse gas concentration increases that have already taken place. The additional warming by 2100 might be anywhere between 0.2 and 0.6°C depending on the precise sensitivity of the climate, with most likely warming of 0.4°C. At the upper end of the scenarios is the A1FI scenario, which yields anywhere from 2.5 to 6.5°C additional warming (with the most likely warming of about 4°C) depending, again, on the sensitivity of the climate. Interestingly, we find that the *scenario uncertainty* and *physical uncertainty* are, in a sense, of nearly the same magnitude. While the most likely warming (i.e., the central estimates for each scenario) ranges from 0.4 to 4°C, i.e., just under a range of 4°C, the range for any one scenario (i.e., A1FI, which ranges from 2.5 to 6.5°C warming) also corresponds roughly to a maximum 4°C range. In this sense, roughly half the spread shown in the various projections of future warming is under our control, i.e., it depends on choices *we* make about future emissions.

There is so much focus on climate projections through 2100 that it is easy to lose sight of the fact that the climate does not magically stop changing at 2100 in the emissions scenarios we have been exploring—indeed, there is, in many cases, significant additional warming and associated changes in climate for several more centuries.

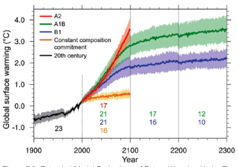


Figure 7.3: Extended Model Projections of Future Warming Under Two IPCC Emissions Scenarios.
Credit: IPCC, 2007

We already discussed that global warming is not predicted to be uniform. High northern latitudes are expected to warm more and faster due, in large part, to the positive ice-albedo feedback, which becomes very strong as Arctic sea ice melts. Land regions are expected to warm faster than ocean regions owing to the ocean's delaying thermal inertia. Some regions will warm more than others, and some may even cool slightly, due to changing atmospheric and oceanic circulation patterns.

Can we see this effects in the actual spatial temperature patterns projected by a state-of-the-art climate model? Let us take a look—we are going to examine the yearly average spatial patterns of surface temperature change in a simulation of the GFDL CM2.1 coupled model (one of the models that contributes to the 20+ member IPCC model ensemble we have been looking at), subjected to the A1B scenario, as it evolves over the entire course of the 21st century.

As you watch the animation below, take note of the overall pattern of warming. Note the latitudinal breakdown of the warming shown to the right of the map. What patterns do you see—are they what you expect? Take note of the variability, both spatially and temporally. Do you see events that resemble El Niño events? Are there any particularly conspicuous, persistent anomalies that emerge over time which you did not expect? You might want to restart the video several times so you can absorb all of the information contained in the animation.

Think About It!

One anomaly you may have noted is the cooling in a small region of the North Atlantic south of Greenland. Any idea what might be responsible for that?

Click for answer.

How much did the model warm in the global mean from 2000 to 2100? How does this compare to the overall spread of projected warming for the A1B scenario shown earlier? If you had to make an educated guess, what "model number" might you suspect this is, looking at the figure comparing cloud radiative forcing for the different IPCC models? Why?

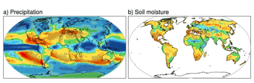
Click for answer.

Surface temperature changes are of course just one of a myriad effects of anthropogenic climate change. Equally if not more important, in terms of its impact on civilization and our environment, are the shifts in rainfall and drought patterns. What do the models have to say about this?

Precipitation and Drought

As we alluded to earlier in the course, climate change is projected to lead to a poleward expansion of the Hadley Cell circulation pattern, which results in an expansion of the zone of subsiding, dry air well out of the subtropics into middle latitudes. This is particularly true in the summer when the ITCZ, jet stream, and polar front shift furthest poleward. As a result, we see decreased rainfall over the large parts of the subtropics through the mid-latitudes, including large parts of North America and Europe. Rainfall increases in the deep tropics where more water vapor is available to be squeezed out of the air and turned into rainfall as it rises within the ITCZ, which migrates north and south of the equator over the course of the year. Large rainfall increases are also seen in sub-polar latitudes owing to a combination of (a) the poleward shift of the jet stream and polar front, bringing this secondary band of rising atmospheric motion further toward the pole and (b) the effect of increased atmospheric moisture, meaning that where it rains, there will be more of it.

In comparing the spatial pattern of precipitation changes (panel *a) Precipitation*) with that of soil moisture (panel *b) Soil moisture*), we might initially be somewhat surprised. The pattern of soil moisture suggests decreases in soil moisture over most of the continents of the world, even in those regions (e.g., the high boreal regions of North America and Eurasia) which see substantial projected *increases* in precipitation. This might seem like a paradox. Is it?



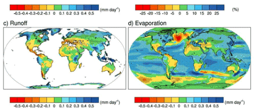


Figure 7.4: Model Projections of Hydrological Changes by end of 21st Century in A1B Emissions Scenario (based on average over all IPCC models). Stippling indicates where there is a consensus among models.
Credit: IPCC, 2007

In fact, there is no paradox here at all. Keep in mind that soil moisture reflects a balance between the water coming in (in the form of precipitation, and runoff) and the water leaving the soil (in the form of evaporation/evapotranspiration). Evaporation is projected to increase over most of the continents, including many regions that are projected to see an increase in precipitation. So, what is actually happening is that even in many areas where rainfall is increasing, soil moisture nonetheless is projected to decrease, and drought thereby worsen, because warmer soils are evaporating water into the atmosphere at a faster rate than water is accumulating at the surface from rainfall or snowfall, even when precipitation is projected to increase. As we will see later, a further complication is the distribution of the rainfall—it is projected to come in fewer, but heavier rainfall events—which, seemingly paradoxically, means that both flooding and drought can become problematic for the very same regions. We will discuss this issue later in our treatment of climate change *impacts*.

Certain projected changes in precipitation are robust, with a fair degree of consensus among models (e.g., much of Canada and Europe). For other regions however (much of the U.S., and much of tropical and subtropical North Africa) there is no clear agreement among models—meaning that the projected changes are highly uncertain. Much of this uncertainty comes from the fact that many of the projected changes in rainfall patterns are related to projected shifts in atmospheric circulation, and these shifts themselves are often quite uncertain. Let us look at this in greater detail.

Atmospheric Circulation Change

We already saw the pattern of projected change in rainfall. It is especially useful to look at this pattern averaged zonally, i.e., by latitude bands, which provides a simpler picture of how rainfall is projected to change as a function of latitude. When we do that, we see a fairly clear latitudinal pattern emerge.

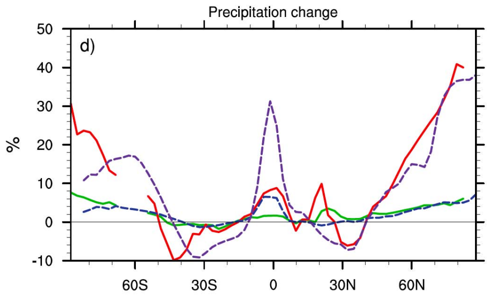


Figure 7.5: Model Projections of Rainfall Changes Over 21st Century in A2 (red and blue) and Constant 2000 CO_2 commitment (purple and green) Scenarios (dashed curves are ocean only; solid curves are land only/ results based on average over all IPCC models).
Credit: IPCC, 2007

Here we see the increase in precipitation near the equator where the ITCZ lies, decreases from the sub-tropics through the mid-latitudes as the Hadley Cell expands poleward, and increases again in sub-polar latitudes where the polar front migrates poleward. In short, we are seeing the effect of the poleward shifting of the zones of rising and descending motion that we reviewed during the overview [57] of atmospheric circulation in our very first lesson.

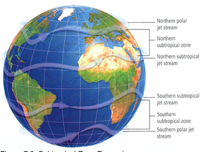


Figure 7.6: Subtropical Zone Expansion.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
© 2015 Pearson Education, Inc.

We first encountered the notion of a potential poleward migration of the jet stream in our previous discussion of the 2003 European heat wave [177], which was a possible harbinger of climate change impacts to come. In this particular case, the sub-tropical jet stream (which lies above the descending limb of the Hadley Circulation in the subtropics, and is associated with the subsidence of warm, dry air in the subtropics) shifted well north of its usual summer location over the northern Sahara desert and southern Mediterranean, well into the middle and even sub-polar latitudes of Europe. That single event encapsulated a pattern that is expected to become more prevalent with future climate change, as the various atmospheric bands, including the Hadley Circulation, and polar front, expand poleward.

It is worth looking more closely, in this context, at one particular metric of the latitudinal shift of the polar front and jet stream, the North Atlantic Oscillation (NAO; we introduced this concept briefly in our discussion of factors influencing Atlantic tropical cyclone activity [178] in Lesson 3). The NAO is a measure of the poleward extent of the Northern Hemisphere jet stream and polar front during northern winter over the North Atlantic ocean. A positive NAO reflects a stronger than usual sub-polar *Icelandic low* surface pressure center and stronger than usual subtropical *Bermuda/Azores high* surface pressure center during the northern winter. It is associated with a strengthened and more northerly storm track. It is associated with warmer and wetter than usual conditions in Europe. By contrast, the negative phase of the NAO is associated with a weaker than normal jet stream, cooler, dryer winters in Europe, and wet winters in the Mediterranean and near/middle east.

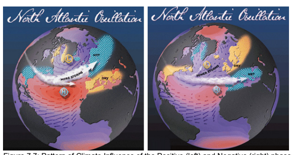


Figure 7.7: Pattern of Climate Influence of the Positive (left) and Negative (right) phases of the NAO.
Credit: Lamont Doherty Earth Observatory [65], Columbia University

Often the NAO is associated with a more hemispherically-symmetric mode of atmospheric variability known as the Arctic Oscillation (AO) or Northern Annular Mode (NAM), which reflects a deeper than usual surface low pressure area throughout the sub-polar belt of the Northern hemisphere, and a deeper than usual surface high-pressure area throughout the sub-tropical belt of the Northern hemisphere. The positive mode of the AO/NAM is associated with a stronger northern hemisphere winter jet stream, while the negative phase is associated with a weakened jet stream.

Climate models project a trend towards a more positive winter NAO/AO/NAM as a result of anthropogenic climate change, due to the changing vertical and latitudinal patterns of temperature (as you may recall from our introductory readings [57], it is the vertical and latitudinal gradients in atmospheric temperatures which drive the jet stream in the first place). This implies a stronger winter jet stream in the Northern Hemisphere (similar changes are projected for the Southern Hemisphere), and stronger surface westerlies in middle latitudes. The stronger westerly winds at the surface further warm winter temperatures over land regions by spreading heat from the relatively warm oceans over the relatively cold continents. A more positive NAO also leads to a relative increase in winter rainfall in the southeastern U.S. and Northern Europe, and dryer conditions in the near and middle east and southern Mediterranean. Given that these regions are currently stressed for their limited available water supply, this projected climate change response could represent a substantial threat to water security in these regions.

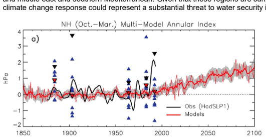


Figure 7.8: Model Projections of the NAO/AO/NAM based on an Average Over all IPCC Models (A1B Emissions Scenario) Compared Against Historical Observations.
Credit: IPCC, 2007

Monsoonal circulation patterns may also change. The most prominent of the monsoons is the South Asian Summer monsoon, which is the source of much of the annual rainfall in heavily-populated regions such as India. The monsoon is driven primarily by the contrast in heating of the oceans and land. The land responds more strongly to summer heating than the oceans, leading to a tendency for a very large-scale thermally-driven circulation cell, similar in some respects to the Hadley Circulation. There is rising motion inland over the Tibetan plateau and sinking motion over the Indian Ocean, so moist warm air over the Indian Ocean is drawn in toward the land, where it rises and condenses out water vapor. This circulation pattern can be influenced by a number of factors, each of which may be altered by anthropogenic climate change. For example, the differential heating of land relative to ocean, projected over the next century, could drive a stronger monsoonal circulation. On the other hand, increased atmospheric stability in South Asia due to latent heating of the atmosphere arising from condensing water vapor in the rising limb of the monsoon could stabilize the vertical temperature profile, inhibiting the Monsoonal circulation. In state of the art models such as those assessed by the IPCC, this latter factor tends to dominate, and the South Asian Summer monsoon is projected to weaken. Seemingly paradoxically, however, the monsoonal precipitation is projected to remain either stable or even increase. In the face of a weakening Monsoonal circulation this has to do with the fact that even though the circulation may be weakened, there will be greater water vapor content in a warmed atmosphere, leading to the potential for greater amounts of rainfall for a given circulation strength. There is still quite a bit of uncertainty on this, however, and a wide spread in projected behavior is seen among the models assessed by the IPCC.

Another pattern of atmospheric circulation that may potentially change as result of anthropogenic climate change is the Walker Circulation, which we discussed in our introduction [179] to the ENSO phenomenon. We will defer any discussion of the uncertain potential changes in this atmospheric circulation pattern to a later section discussing potential change in modes of atmospheric-ocean variability.

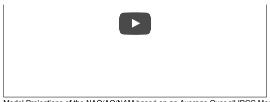
Of course, it is not only the atmosphere which is projected to change in its circulation patterns as a result of anthropogenic climate change, but also the ocean.

Oceanic Circulation Changes

By far, the most critical issue regarding climate change impacts on ocean circulation patterns involves the thermohaline/conveyor belt /meridional overturning circulation which we discussed earlier in the course (e.g., Ocean Circulation page of Lesson 1 [180] and the Oceans page of Lesson 3 [111]). As we discussed earlier, it was once thought that global warming could paradoxically lead to cooling over a wide region of the globe by sending large amounts of fresh water into the high-latitudes of the North Atlantic, where it would freshen the surface waters and inhibit the formation of dense surface waters whose sinking in the sub-polar North Atlantic constitutes the descending limb of the conveyor belt circulation. Since this ocean current system is a substantial contributor to the transport of heat to the high latitudes of the North Atlantic, such an occurrence could lead to widespread cooling of the North Atlantic and neighboring continental regions. Indeed, scientists believe this happened during the Younger Dryas event [150] toward the end of the last ice age, between 13,000 and 12,000 years ago as the climate was warming during the initial phase of deglaciation.

Of course, as noted during our earlier discussion of the Younger Dryas event, the two situations are quite different in many respects. Toward the end of the last ice age, there was much ice to melt, and far greater potential to flood the North Atlantic with extremely large amounts of fresh water. Today, however, the ice sheets are much diminished, and there is less snow and ice available to melt. Nonetheless, certain simple climate models, such as the CLIMBER model used by the Potsdam Institute for Climate Change Impacts, suggest that global warming could lead to a substantial weakening of the thermohaline circulation and a fairly dramatic cooling of the North Atlantic and neighboring regions.

Model Projections



Model Projections of the NAO/AO/NAM based on an Average Over all IPCC Models (A1B Emissions Scenario) Compared Against Historical Observations.
Credit: Potsdam Institute for Climate Impacts Research [181]

We have seen that current state-of-the-art climate models show only a weak semblance of this effect, however. Earlier in this lesson [182], we saw that the GFDL CM2.1 coupled model produces a very moderate cooling over a small region of the North Atlantic in response to anthropogenic warming, in part due to a slight weakening of the thermohaline circulation due to the mechanisms discussed before.

In fact, this model exhibits a larger response than most of the 20+ models used in the most recent IPCC assessment. If we look at the average pattern of surface temperature change over the next century in the A1B scenario, averaged over all 20+ models, we see only the vaguest hint of the effect. Nowhere in the North Atlantic do we see any cooling. Instead, we see a small region in the North Atlantic, south of Greenland and Iceland, where there is less warming than in most other regions. Western Europe also sees somewhat less warming than most other regions due to the downstream effects. So if we take the composite response of the IPCC models as our best current assessment, we are led to conclude that the very, very small grain of truth to the *Day After Tomorrow* [104] scenario, is that global warming may lead to a bit less warming in some parts of the North Atlantic and neighboring regions, owing to a slowing of the oceanic thermohaline circulation.

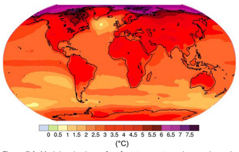


Figure 7.9: Model projections of surface temperature changes by end of 21st Century in A1B Emissions Scenario (based on average over all IPCC models).
Credit: IPCC, 2007

If we look at a more direct measure of the thermohaline circulation in the models, namely the intensity of the northward current associated with the meridional overturning component of the ocean circulation and thermohaline circulation, we see that in very few models does it actually collapse. In some models, it does weaken substantially, but in most models, it weakens only very modestly; and in some models, it marches along at near its current intensity, as if nothing happened at all! Why is this? More elaborate climate models, which contain detailed three-dimensional representations of ocean components that finely resolve boundary currents and even the eddies in these currents, tend to show far more robustness of the MOC than the simpler models, which ignore lateral ocean currents, eddies, etc. The state-of-the-art models have more *degrees of freedom*, i.e., more ways to circulate and transport water, heat, and salinity, and it is, therefore, much harder to cut off the poleward transport of heat by the ocean circulation, as there are multiple components of the oceanic circulation that can deliver heat poleward. These models typically, for related reasons, have a far more stable thermohaline circulation than simpler, earlier ocean models.

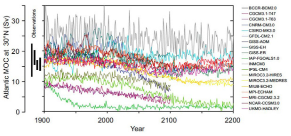


Figure 7.10: Model projections of the strength of the meridional overturning circulation (MOC) of the North Atlantic (measured in "Sverdrups" which is a million cubic meters of water transported poleward per second) in A1B Emissions Scenario (based on average over all IPCC models).
Credit: IPCC, 2007

Models of Climate Variability

We know that the ENSO phenomenon has a profound influence on climate on inter-annual timescales, leading to substantial regional alterations in temperature and rainfall patterns around the world, and influencing important phenomena, such as Atlantic hurricane activity. Needless to say, one key question of climate change is how the characteristics of ENSO might change in the future.

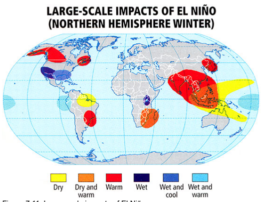


Figure 7.11: Large-scale impacts of El Niño.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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One of the great limitations in projecting future regional climate change is that we cannot yet confidently assess how ENSO will change in the future. The state-of-the-art models used in the most recent IPCC assessment do not show any clear consensus regarding how anthropogenic climate change will influence the characteristics of ENSO.

Most models (the models that fall in the right half of the plot below) project an overall more El Niño-like climate, that is to say, a weakened *Walker Circulation*, weakened trade winds and oceanic upwelling in the eastern and central tropical Pacific, and a surface temperature pattern wherein the eastern equatorial Pacific warms up more than the central and western equatorial Pacific ocean. However, a few state-of-the-art climate models (i.e., the three that fall in the left hand of the plot below) project the opposite, a more La Niña-like pattern. One might be tempted to go with the majority, the El Niño-trending models. However, paleoclimate evidence of the response of ENSO to past natural changes in radiative forcing (including research your course author has been involved in [183]) suggests the possibility that the minority of models (i.e., the La Niña-trending models) might be right. This issue has implications for how anthropogenic climate change might influence Atlantic hurricane activity (this is also an issue that your course author has done research on [184]). We will discuss climate change influences on tropical cyclone activity in more detail in our next lesson.

There is even greater uncertainty regarding the amplitude of ENSO variability, i.e., whether individual El Niño and La Niña events will become larger (models that fall in the upper half of the plot below) or smaller (models that fall in the lower half), with an equal split between the IPCC models with regard to which of the two possiblities is more likely. Since larger El Niño events have a greater impact on regional weather patterns, while smaller events have a lesser impact, this issue has significant implications for projected climate change impacts.

The large uncertainties in projecting changes in ENSO likely result from a combination of factors, including the tendency for state-of-the-art climate models to produce an unrealistic *split ITCZ* in the eastern tropical Pacific, which biases the strength of the trade winds and equatorial upwelling; the still course resolution of the oceanic component of the models, which may lead to inaccuracies in the ocean wave disturbances that are important for El Niño and La Niña; and uncertainties in the behavior of marine stratocumulus clouds, which play an important role in equatorial Pacific radiation and heat budget.

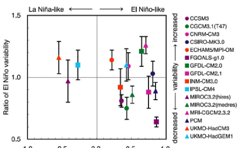


Figure 7.12: Model Projections of Changes in ENSO in Response to Doubling of CO_2 Concentrations.
Credit: IPCC, 2007

Lesson 7 Summary

In this lesson, we examined some of the key anthropogenic climate change projections. Some of the main findings were:

- climate models project anywhere from 0.2 to 7°C warming over the next century, depending on two critical variables: (1) what decisions society makes regarding future carbon emissions and (2) currently irreducible uncertainties regarding the sensitivity of the climate to greenhouse gas radiative forcing;
- the primary source of spread in projected warming is factor #1, the uncertainty in future emissions. Were we to freeze greenhouse gases at their current levels, the average projection among models is an additional warming of 0.5°C. This scenario is highly unlikely, as it is very difficult to find a pathway to zero emissions in the near future. On the other hand, were we to pursue *business-as-usual* emissions scenario (e.g., continue on the A1FI scenario), the average projection among models is for an additional warming of 4°C;
- the impact of factor #2, the uncertainty in climate sensitivity, is nonetheless quite significant. In the A1FI scenario, the globe could warm anywhere from a lower bound of 2.5°C to an upper bound of 6.5°C, depending on which particular climate model is used;
- the variability in temperatures in both space and time is projected to be considerable. High latitudes warm more than low latitudes owing to positive feedbacks related to the melting of ice, and land warms more than oceans due in large part to the greater thermal inertia of the oceans. Even as the globe warms, there will continue to be cold periods over particular regions related to ENSO and other sources of natural variability;
- precipitation is projected to increase in the tropics and sub-polar latitudes, while decreases are projected for sub-tropical through mid-latitude regions. These changes reflect a combination of the effects of shifting storm tracks and the potential for a warmer atmosphere to hold more water vapor;
- continental drought becomes more widespread over much of the continents. This results from a tendency for increased evaporation to dry out soil, even in many regions that see an increase in precipitation;
- anthropogenic climate change leads to substantial changes in atmospheric circulation, including a poleward shift of the descending branch of the Hadley Circulation and of the jet streams, polar front, and storm tracks. These changes also include a weakening of monsoonal circulations, and possible, but uncertain, impacts on the Walker Circulation pattern associated with ENSO;
- the NAO/AO/NAM mode of variability, tied with variations in the position of the Northern Hemisphere winter jet stream, is projected to become more positive, associated with a northward displaced storm track and warmer, wetter winter conditions in regions such as Europe, but drier conditions in semi-arid regions such as the Mediterranean and Near and Middle East;
- a modest weakening is projected for the meridional overturning ocean circulation (variously referred to as the thermohaline circulation and conveyor belt circulation). State-of-the-art models, however, project a far weaker effect than what was once considered possible; rather than leading to cold conditions over the North Atlantic and neighboring regions, what is currently projected is only a moderate decrease of the warming in a small region in the North Atlantic ocean south of Greenland;
- projected changes in the character of the El Niño/Southern Oscillation are uncertain. Model projections are divided with respect to whether the future climate state will be more like El Niño or La Niña, and whether individual El Niño and La Niña events will be larger or smaller.

Reminder - Complete all of the lesson tasks!

You have finished Lesson 7. Double-check the list of requirements on the first page of this lesson [176] to make sure you have completed all of the activities listed there before beginning the next lesson.

Lesson 8 - Projected Climate Changes, part 2

The links below provide an outline of the material for this lesson. Be sure to carefully read through the entire lesson before returning to Canvas to submit your assignments.

Introduction

About Lesson 8

We will now look at some of the more subtle, and indeed, less certain, future climate changes projected by climate models. These include changes in the cryosphere—that is glaciers, ice sheets, and sea ice; changes in sea level, which reflect an integration of a number of factors including melting ice and warming oceans; and changes in extreme weather events including tropical cyclones. While it is more difficult to accurately project changes in these attributes of the climate system, their profound potential impact on civilization and our environment necessitate their consideration.

What will we learn in Lesson 8?

By the end of Lesson 8, you should be able to:

- qualitatively assess the potential impacts of increasing greenhouse gas concentrations on the distribution of ice on the Earth's surface, global sea level, tropical storm activity, severe weather, and other relevant climatic and meteorological phenomena;
- discuss the potential importance of carbon cycle feedbacks; and
- discuss the concept of climate tipping points, and their potential impacts.

What will be due for Lesson 8?

Please refer to the Syllabus for specific time frames and due dates.

The following is an overview of the *required activities* for Lesson 8. Detailed directions and submission instructions are located within this lesson.

- Participate in Lesson 8 discussion forum: Climate Change Projections.
- Read:
  - IPCC Fifth Assessment Report, Working Group 1 [2]
  - Summary for Policy Makers [3], Future Global and Regional Climate Change
    - E.4 Ocean: p. 24
    - E.5 Cryosphere: p. 24-25
    - E.6 Sea Level: p. 25-26
    - E.7 Carbon and Other Biogeochemical Cycles: p. 26-27
  - Dire Predictions, v.2: p. 110-117

Questions?

If you have any questions, please post them to our *Questions?* discussion forum (not e-mail), located under the Home tab in Canvas. The instructor will check that discussion forum daily to respond. Also, please feel free to post your own responses if you can help with any of the posted questions.

Sea Ice, Glaciers, Ice Sheets

Critics often argue that climate scientists are alarmist and overstate projections of future climate change (if you have any doubt about this, just do a Google news search [185] on e.g., "warming" and "alarmist" and see what turns up—warning: the Internet is a wild frontier of information, misinformation, and outright disinformation—you should always question the objectivity and qualifications of any apparent news sources).

Ironically, in many cases, the climate science community has been overly cautious and conservative in their projections (see for example this review by Rahmstorf et al. in the journal *Science* [186]), perhaps, in part, out of fear of being labeled as alarmist by its detractors). One case in point is the decline in Arctic summer sea ice extent. The decline in minimum summer Arctic sea ice has outpaced even the most dramatic of the IPCC projections in recent decades. One possibility is that there is missing physics involved—related, for example, to the mechanics of ice fracturing and deformation, that is causing the models to underpredict the fragility of sea ice and the possible feedbacks involved.

Following the dramatic decline of summer 2007 (the minimum value in the red curve below), there was even fear that we had crossed a previously unknown *tipping point* from which Arctic sea ice would not be able to recover (we will have far more to say about possible climate *tipping points* [187]later in this lesson). If so, that would have dire implications for, e.g., animal species such as the polar bear, which require sea ice environment for their existence. Recent scientific work suggests that there is no such tipping point, however, and that the environment of the polar bear can be preserved given ongoing efforts aimed at mitigating future climate change (see this news account [188] from the Churchill "Plight of The Polar Bear" video [189] introduced in our first lesson—you may recognize the lead scientist on the study, Steven Amstrup,). Those efforts would not be easy, however; the study finds that future warming would have to be kept below 1.25°C, which corresponds to roughly 2°C warming relative to pre-industrial time. As we have seen before, such a target would likely require stabilizing CO_2 concentrations at 450 ppm or lower.

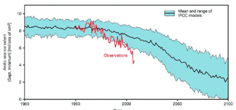


Figure 8.1: Observed Trend in Summer Minimum Arctic Sea Ice Extent Compared Against IPCC Model Historical Simulations and Projections.
Credit: The Copenhagen Diagnosis [190]

One problem that declining sea ice does not contribute to (at least, not to any practical extent) is the problem of global sea level rise. The melting of land ice, nonetheless, will contribute to sea level rise. As we saw in a previous lesson, mountain glaciers and ice caps around the word have undergone substantial retreat [191] over the past century. They are projected to contribute a fraction of a meter of sea level rise over the next century (for example, see this news article about a study recently published in the journal *Nature Geoscience* [192]).

The melting of the major continental ice sheets will also contribute to future sea level rise. This is part of the reason that so much attention is paid to the stability of the two major ice sheets: the Antarctic and Greenland ice sheets. In a previous lesson, we already saw that the two ice sheets appear to have entered into a regime of negative mass balance, i.e., they are now losing ice [191]. Let us take a look at the best available projections for what is likely to happen to these ice sheets under a global warming scenario.

Shown below is a simulation of the process of Antarctic sea ice retreat in response to global warming. The simulation was done by one of the leading ice sheet modelers, Penn State's own David Pollard [193]. As we see here, it is primarily the low elevation West Antarctic half of the Ice Sheet that is prone to melting, and the process of collapse, at least in this simulation, is potentially quite slow, occurring on the millennial timescale.

Simulation of collapse of West Antarctic ice sheet in global warming scenario.
Credit: Penn State Earth System Science Center [194]

A similar finding is obtained for the Greenland ice sheet. Shown below is a simulation using NCAR climate model coupled to a dynamical model of the Greenland ice sheet. This particular simulation suggests that it might take a millennium or longer for substantial loss of Greenland ice due to projected global warming.

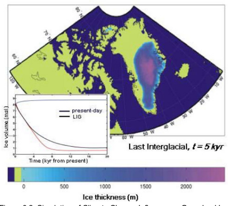
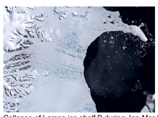


Figure 8.2: Simulation of Climate Change Influence on Greenland Ice Sheet Based on NCAR Climate Model (the red curve Indicates the response to an anthropogenic quadrupling of CO2 concentrations relative to pre-industrial levels).
Credit: National Center for Atmospheric Research [195]

Yet, processes discussed on the Sea Ice page of Lesson 3 [191] could lead to ice sheet collapse on timescales much faster than suggested by these model simulations. The formation of fissures known as *moulins* allow meltwater to penetrate to the bottom of the ice sheet, where it lubricates the base of the ice sheet, allowing ice to be more easily exported through ice streams out into the ocean. The physics responsible for this phenomenon are not well represented in current ice sheet models.

Figure 8.3: A moulin that formed in the Greenland ice sheet.
Credit: Roger Braithwaite, UNEP

Another effect that is not well represented by the models is the *buttressing* effect provided by ice shelves. Ice shelves provide support to the interior ice sheet through an effect not unlike that of the flying buttress [196] employed so widely in medieval architecture. As the ice shelves themselves disintegrate, there is the potential for destabilization of the inland ice, which is then free to break up and calve into the ocean. This effect was well documented during the disintegration of the Larsen B ice shelf of the Antarctic Peninsula in January-March (austral summer) 2002. The ice shelf, roughly the size of the state of the Rhode Island, disintegrated in little over a month. In the months following the breakup of the ice shelf, an acceleration in the streaming of inland ice to the ocean was observed, suggesting the possibility that such dynamical processes could accelerate the collapse of the West Antarctic ice sheet—the part of the Antarctic Ice Sheet most susceptible to break-up as a rule of anthropogenic warming. An even more recent, though less dramatic, example is discussed here. [197] Once again, the observation of such processes in nature, and the fact that the ice sheets already appear to be losing ice mass [191], suggests that the breakup of continental ice shelves could proceed considerably faster than is suggested by current generation ice sheet models.



Collapse of Larsen ice shelf B during Jan-Mar 2002.
Credit: National Snow and Ice Data Center [198]

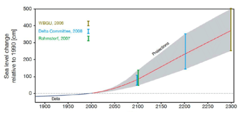
Regardless of the precise timescale of decay of the continental ice sheets, it is quite possible, as discussed in the "Greenland ice sheet faces 'tipping point in 10 years' " [199] article by Richard Alley [200]—another of the world's leading ice sheet experts also here at Penn State—that we might commit to the melting of the Greenland ice sheet much sooner than the models project—in as soon as a decade—by reaching a critical level of greenhouse gases in the atmosphere whereby we commit to the initial warming necessary to set in motion the sequence of positive feedbacks leading to the ultimate destruction of the ice sheet. Once again, we encounter the notion of a *tipping point*—a notion to be explored later in this lesson.

Sea Level Change

There are several components involved in projecting future sea level. One—the thermal expansion of the oceans, is fairly straightforward, and the only real uncertainty involved with that component is the warming itself, and the rate at which it is mixed down beneath the ocean surface. This contribution is projected to be modest, amounting to only a fraction of a meter over the next century. The 2nd component is the contribution from melting mountain glaciers and ice caps. This contribution, too, is likely to be modest, only a small fraction of a meter within the next century. The 3rd contribution—that from the melting of the two major ice sheets—is both the largest and the most uncertain.

As alluded to earlier, even modest additional warming could set in motion the collapse of all or most of the Greenland Ice Sheet (which would add 5-7 meters of sea level rise) and the west Antarctic Ice Sheet (which would add roughly another 5 meters). In our discussion of the projected changes in the Greenland and Antarctic Ice Sheet in the previous section, we saw that there are significant uncertainties regarding the timescale of this disintegration, however. Current ice sheet models suggest that the collapse of the ice sheets may take many centuries. Yet we know there are reasons to be skeptical about current ice sheet models. They are missing some physics that appears to be important in the real world—i.e., the physics responsible for the formation of moulins, and the possible lost of buttressing support by decaying ice shelves—that could allow for much faster disintegration. Indeed, the very same ice sheet models that predict a slow, multi-century breakup of the ice sheets did not predict that ice sheet loss would be observed for many decades, and yet, as we have seen that this loss already appears to be underway [191].

The uncertainties in projecting the ice sheet melt obviously complicate projections of future sea level rise, since this is potentially the largest contributor to global sea level rise. In the IPCC AR4 report from 2007, the IPCC simply neglected this contribution because they considered it too uncertain to estimate. This means that that their formal projections were almost certainly an absolute lowest estimate. (However, the IPCC AR5 report does include an estimate of the ice sheet contribution, leading to an increased sea level change projection of about 0.60 m by 2100 under SRES A1B.) Is it possible to provide a more realistic estimate? German climate scientist Stefan Rahmstorf [201] and collaborators have used an alternative, so-called semi-empirical approach to projecting future sea level rise. This approach uses the historical relationship between past changes in sea level rise and global mean temperature to construct a statistical model, and, in principle, incorporates the contribution from melting ice sheets, though obviously with some amount of uncertainty. The semi-empirical projections suggest the possibility of more than a meter (roughly 3 feet) of sea level rise by 2100, and more than 3 meters (roughly 9 feet) of sea level rise by 2200. We will look at the likely impacts of such amounts of sea level rise when we focus on climate change impacts later on in the course.



Year

Figure 8.4: Projections of future global sea level rise ranging over the various IPCC scenarios, based on semi-empirical projections.
Credit: The Copenhagen Diagnosis [190]

This example also provides a nice introduction to the concept of semi-empirical models. We will examine a similar semi-empirical modeling approach in our discussion of projected changes in Atlantic tropical cyclone activity in the next section.

Tropical Cyclones / Hurricanes

Like global sea level rise, climate change impacts on tropical cyclone activity could have profound societal and environmental implications. And, as with global sea level rise, impacts of tropical cyclone activity represent a substantial challenge scientifically, as there are many uncertainties that are involved. Tropical cyclones occur at spatial scales that are not well resolved by current generation climate models. Various approaches have been used to try to get around this problem, and you can find a discussion by your course author of the relative merits of these different approaches on the RealClimate website [202].

One approach has been to take a finer-scale atmospheric model that is capable of producing tropical cyclone or at least tropical cyclone-like disturbances, and nest it within a larger-scale climate model. The large-scale boundary conditions from the climate model simulation are then used to drive the storm-resolving model. However, the storm-resolving models used do not generally resolve the critical inner core region of the tropical cyclone, and the cyclones produced are not especially realistic. The models, for these reasons, cannot reproduce major hurricanes (category 4 or 5), and, for this reason, one might call into question their ability to capture changes in hurricane behavior associated with climate change.

An alternative approach used by hurricane scientist Kerry Emanuel [63] of MIT employs embedded modeling. Small-scale disturbances, similar to those that generate real-world tropical cyclones, are randomly distributed in a climate model in a way that mimics the distribution of real world disturbances. Some of these disturbances will find themselves in a favorable environment, others will not—that is determined by the large-scale climate as represented by the climate model. A finely resolved dynamical model, that does indeed resolve the key inner core structure of a tropical storm, is embedded within the climate model and is used to track each such disturbance and model any potential development and intensification. Using this approach, Emanuel is able to closely match the observed tropical cyclone climatologies (i.e., the numbers and intensities of tropical cyclones) in the various basins of tropical cyclone activity. He is also able to reproduce the observed trend, including the increase in tropical cyclone numbers and intensities in the Atlantic in recent decades, when he uses the actual large-scale observations—in place of a climate model simulation—to drive his model of tropical cyclone genesis and intensification.

So, what results does this approach yield when driven with projections of future climate change? The results of the analysis are shown below. What we see, first of all, is that there is quite a bit of variability in the results you get, depending on (a) which particular climate model projection is used and (b) which particular tropical cyclone-producing basin you are looking at. Globally, the number of tropical cyclones may actually *decrease*, but the power dissipation and intensity are projected to *increase* globally. We are particularly interested in Atlantic tropical cyclones and hurricanes, since they pose the greatest threat as far as North American impacts are concerned.

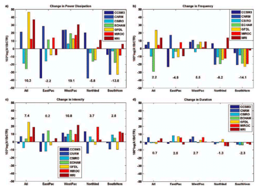


Figure 8.5: Projected changes in Tropical Cyclone Characteristics in various basins over the next two centuries based on the A1B projections based on an embedded modeling approach.
Credit: Emanuel et al, Hurricanes and Global Warming, Bulletin of the American Meteorological Society, 347-367, DOI:10.1175/BAMS-89-3-347, 2008.

We see that majority of the models yield a substantial increase in power/intensity of Atlantic tropical cyclones. A majority also indicate an increase in the number of Atlantic tropical cyclones, but this is highly variable, with at least one of the seven models examined indicating a substantial decrease. The divergence among the various projections of the different models reflects the competition between factors favoring increased power and activity—e.g., warmer oceans and greater energy for driving storms—and other factors, such as changes in atmospheric conditions influencing, e.g., vertical wind shear, that are tied to dynamical changes in the climate system. In particular, the uncertainty in projecting changes in ENSO dynamics (from Lesson 7) [203] comes into play here. El Niño events are associated with increased wind shear over key regions of the Caribbean and tropical Atlantic where tropical cyclone tend to form, so a more El Niño-like climate would tend to mitigate any increases in Atlantic tropical cyclone activity. By contrast, a more La Niña-like climate would mean an even more favorable atmospheric environment. So, the current uncertainty in projecting changes in ENSO and the Walker Circulation translate to uncertainties in projecting future changes in Atlantic tropical cyclone activity.

Yet, an entirely different approach to projecting changes in tropical cyclone activity involves the use of a *semi-empirical approach* similar in spirit to the semi-empirical modeling approach we encountered in our study of sea level rise projections in the previous section. The semi-empirical approach involves, once again, using a statistical model to relate the phenomenon of interest (tropical cyclone activity) to the climate factors we know appear to govern year-to-year changes in activity today (e.g., tropical Atlantic SSTs, El Niño, and the North Atlantic Oscillation). In fact, you may recall that this is the very same statistical model that you constructed yourself in problem set #2 [68] using these three factors as predictors of Atlantic annual tropical cyclone numbers in a multivariate regression. That approach has been validated, to some extent, by comparisons of predictions of pre-historical changes in Atlantic tropical cyclone activity with geological records of Atlantic tropical cyclone activity spanning the past millennium. You can hear about some work the course author has done in this area in this NSF video press conference [204].

We are now going to use the very same statistical model you developed in problem set #2 to project future changes in Atlantic tropical cyclone numbers. You can vary the predicted tropical Atlantic warming (over the rough range of IPCC projected changes in global mean temperature over the next century), and you can vary the scenario for how ENSO will change (anywhere from a substantial trend toward an El Niño-like state, i.e., positive Niño3 index values, to a substantial trend toward a La Niña-like state, i.e., negative Niño3 index values). For simplicity, we will assume that the NAO remains fixed at its modern day average (its role in the statistical model is pretty minute anyway). Applying the regression coefficients of the statistical model to these climate change projections then yields a semi-empirical projection of future annual Atlantic tropical cyclone numbers. The flash application is provided below—please play around with the two dropdown boxes in the lower part of the figure below and get a sense of how the range of uncertainty in the climate change projections translates to uncertainty in projected future tropical cyclone activity.

Be prepared to discuss your findings in this week's discussion forum.

Extreme Weather

We saw earlier in the course [177] that climate change already appears to have influenced the frequency and intensity of various types of extreme weather events. The observed warming so far amounts to less than a 1°C relative to pre-industrial time. Given projected warming of several more degrees C over the next century (depending on the precise emissions scenario), the future increases in extreme weather events can be expected to be far larger than what we have observed thus far.

A large increase in the incidence of extreme precipitation events is expected. As we know, warmer oceans evaporate water into the atmosphere at a faster rate, and a warmer atmosphere can hold more water vapor. These features imply a more vigorous hydrological cycle, and heavier individual events when conditions are conducive to precipitation. Basic atmospheric physics tells us that 1°C warming we have seen already implies roughly 2% higher concentration of water vapor in the atmosphere on average, and correspondingly, roughly 2% more precipitable water during any particular precipitation event. Depending on the particular emissions scenario, we can expect a several fold larger increase over the next century. Since flooding is associated with large accumulations of rainfall over short periods of time, this increase in precipitation intensity implies greater potential for flood conditions—ironically, even for regions that on average see greater drought, i.e., more dry days (something we alluded in our discussion of general precipitation trends [205] in the previous lesson).

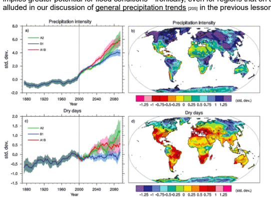


Figure 8.6: Model Projections of Changes in Precipitation Intensity (top) and Frequency of Dry Days (bottom) by end of 21st Century in Various Emissions Scenario (based on average over all IPCC models).
Credit: IPCC, 2007

Where atmospheric temperatures are above freezing, we expect precipitation to fall as rain, but where temperatures are below freezing, we expect it to fall as snow. Climate change deniers often claim [206] that heavy winter snowfalls argue against the reality of global warming. Nothing could be further from the truth, however. In any realistic scenario of the next century, we will still have winter. It will still be cold enough for snow over large parts of North America in winter, and as the atmosphere continues to warm and hold more water vapor, we expect those snowfalls to be heavier. In fact, there is a bit of a double whammy involved here: we also know that mid-latitude winter storms will likely become more intense—something we already see evidence of in meteorological observations [177]. These more intense storms will be associated with stronger frontal boundaries, further increasing the potential for large snowfalls (you can find your course author commenting on this issue [207] here).

Profound changes are, of course, also expected in temperature extremes. Heat waves, which, as we saw earlier in the course, have already increased [177] in duration and intensity owing to the warming of the past century, are projected to be subject to further increases over the next century, with the details depending, of course, on the emissions pathway. Not surprisingly, extreme cold days are projected to decrease in number.

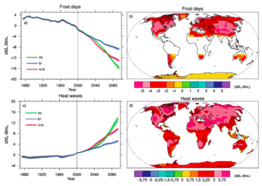


Figure 8.7: Model Projections of Changes in Frequency of Frost Days (top) and Heat Waves (bottom) by end of 21st Century in Various Emissions Scenario (based on average over all IPCC models).
Credit: IPCC, 2007

In our original discussion of extreme weather in Lesson 3 [177], we likened the incidence of extreme weather to the rolling of a die. Sixes come up relatively rarely in the roll of a *fair* die (1 in 6 rolls to be precise). Let us think of an unusually warm summer day (say, a 90F day in State College PA in July) as a rolling of a six. We have seen that the incidence of extreme warm days in the U.S. has roughly doubled since the mid 20th century. Think of that as a doubling of the probability of rolling a six. You got a sense for how apparent such a change in the odds might be in the day-to-day weather variations by playing a game where you compared a fair die and one that had been loaded to double the probability of sixes. It took a fair number of rolls in that case [177]—perhaps as many as 10 or so—to be pretty confident from the pattern of rolls as to which one was the loaded die. Now, we will *double* the probability once more—so there is now a *four-fold increase* in the likelihood of rolling a six in the loaded die relative to a fair die. Using the updated version of the dice rolling application below, see how many rolls of the die it takes you *now* to figure out which one is loaded and which one is fair. What does this tell you about the degree to which the "loading of the die" will become more apparent in observations of extreme weather as time goes on?

Be prepared to discuss your observations in this week's discussion.

Carbon Cycle Feedbacks

As we saw earlier in the course, the airborne fraction of CO_2 in the atmosphere has increased by only half as much as it should have given the emissions we have added through fossil fuel burning and deforestation. We know that CO_2 must be going somewhere.

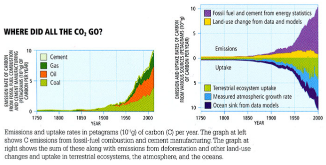


Figure 8.8: Annual change in atmospheric CO_2 concentrations.
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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Indeed, it is being absorbed by various reservoirs that exist within the global carbon cycle. As we saw earlier, in Lesson 1 [127], only 55% of the emitted carbon has shown up in the atmosphere, while roughly 30-35% appears to be going into the oceans, and 15-20% into the terrestrial biosphere.

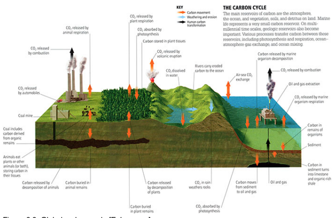

Figure 8.9: Global carbon cycle.[Enlarge [208]]
Credit: Mann & Kump, *Dire Predictions: Understanding Climate Change, 2nd Edition*
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The problem is that this pattern of behavior may not continue. There is no guarantee that the ocean and terrestrial biosphere will continue to be able to absorb this same fraction of carbon emissions as time goes on, and that leads us into a discussion of so-called *carbon cycle feedbacks*.

If we consider the oceans, for example, there are a number of factors that could lead to decreased uptake of carbon as time goes on. Like a warm can of Coke, which loses its carbonation when you warm it up and remove the top, the ocean's CO_2 solubility decreases as the ocean warms. When we look at the pattern of carbon uptake in the upper ocean, we see that one of the primary regions of uptake is the North Atlantic. This is, in part, due to the formation of carbon-burying deep water in the region. In a scenario we have explored in Lesson 7 [209], the North Atlantic overturning circulation could weaken in the future (though as we have seen, there is quite a bit of uncertainty regarding the magnitude and time frame of this weakening). If that were to happen, it would eliminate one of the ocean's key carbon-burying mechanisms, and allow CO_2 to accumulate faster in the atmosphere. On the other hand, the biological productivity of the upwelling zone of the cold tongue region of the eastern and central equatorial Pacific is a *net source* of carbon to the atmosphere, *from* the ocean. More El Niño-like conditions in the future could suppress this source of carbon, but more La Niña-like conditions could increase this source, further accelerating the buildup of CO_2 in the atmosphere. So uncertainties in the future course of oceanic uptake abound, but, on balance, it is likely that this uptake will decrease over time, yielding a *positive carbon cycle feedback*.

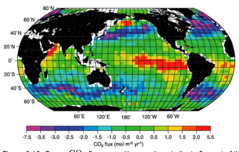


Figure 8.10: Ocean CO_2 fluxes: positive numbers indicate flux out of the ocean.
Credit: IPCC, 2007

Other ocean carbon cycle feedbacks relate to the phenomenon of *ocean acidification*, which results from the fact that increasing atmospheric CO_2 leads to increased dissolved bicarbonate ion in the ocean (a phenomenon will discuss further in our next lesson on climate change impacts). On the one hand, this process interferes with the productivity of calcite-skeleton forming ocean organisms, such as zooplankton, which bury their calcium carbonate skeletons on the sea floor when they die. This so-called *oceanic carbon pump*, is a key mechanism by which the ocean buries carbon absorbed from the atmosphere on long timescales. So any decrease in the effectiveness of the ocean's carbon pump would represent a positive carbon cycle feedback. On the other hand, since calcifying organisms release CO_2 into the water as they build their carbonate skeletons, a decrease in calcite production by these organisms will reduce CO_2, amounting to a negative carbon cycle feedback.

There are a number of other carbon cycle feedbacks that apply to the *terrestrial biosphere*. They vary anywhere from a strong negative to a strong positive feedback. Among them are (a) warmer land increasing microbial activity in soils, which releases CO_2 (a small positive feedback), (b) increased plant productivity due to higher CO_2 levels (a strong negative feedback). Finally, there is the negative silicate rock weathering feedback which we know to be a very important regulator of atmospheric CO_2 levels on very long, geological timescales: a warmer climate, with its more vigorous hydrological cycle, leads to increased physical and chemical *weathering* (the process of taking CO_2 out of the atmosphere by reacting it with rocks), through the formation of carbonic acid, which dissolves silicate rocks, producing dissolved salts that run off through river systems, eventually reaching the oceans.

While each of these potential carbon cycle feedbacks are uncertain in magnitude—and even in sign in some cases (see the various coloured bars in the figure below), the net result of all of these feedbacks appears to be a *net positive carbon cycle feedback* (the black bar shown).

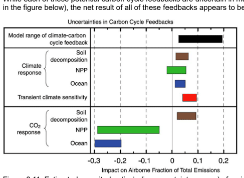


Figure 8.11: Estimated magnitudes (including uncertainty ranges) of various potential oceanic and terrestrial carbon cycle feedbacks, expressed in terms of positive or negative estimated change in the airborne fraction of CO_2 (based on average net increase by 2100 among the various climate models).
Credit: IPCC, 2007

Other potential positive carbon cycle feedbacks that are even more uncertain, but could be quite sizeable in magnitude, are methane feedbacks, related to the possible release of frozen methane currently trapped in thawing Arctic permafrost, and so-called "clathrate"—a crystalline form of methane that is found in abundance along the continental shelves of the oceans, which could be destablized by modest ocean warming. Since methane is a very potent greenhouse gas, such releases of potentially large amounts of methane into the atmosphere could further amplify greenhouse warming and associated climate changes.

The key potential implication of a net positive carbon cycle feedback is that current projections of future warming such as those we have explored previously in Lesson 7 [182], may actually underestimate the degree of warming expected from a particular carbon emissions pathway. This is because the assumed relationship [170] between *carbon emissions and CO_2 concentrations* would underestimate the actual resulting CO_2 concentrations because they assume a fixed airborne fraction of emitted CO_2, when, it fact, that fraction would instead be increasing over time. While the magnitude of this effect is uncertain, the best estimates suggest an additional 20-30 ppm of CO_2 per degree C warming, leading to an additional warming of anywhere from 0.1°C to 1.5°C relative to the nominal temperature projections [182] shown in earlier lessons.

Earth System Sensitivity

As we saw in the previous section on carbon cycle feedbacks, there are some limitations in the traditional framework for assessing the climate response to anthropogenic forcing. In the case of carbon cycle feedbacks, the assumptions implicit in that framework regarding the relationship between carbon emissions and resulting CO_2 concentrations may underestimate the future increase in CO_2 levels, and the degree of climate change.

Another problem in the traditional framework is that the assessment of *climate sensitivity*—a topic we have looked at in depth in Lesson 5 of this course [210]—may be too limited. Implicit in the traditional definition of climate sensitivity is the so-called "Charney" notion of climate sensitivity—named after a famous climate scientist (Jule Charney [211]) and a late 1970s report of the National Academy of Sciences known as the "Charney Report" [212] which provided the first real estimate of the range of uncertainty in the equilibrium climate sensitivity. The concept of climate sensitivity described in this report, sometimes called the "Charney Sensitivity", envisions the equilibrium sensitivity of Earth's climate to CO_2 forcing as the equilibrium response of the climate system to a doubling of CO_2 concentrations including all *fast feedbacks*—that includes changes in water vapor, clouds, sea ice, and perhaps even small ice caps and glaciers.

The limitation implicit in this definition becomes apparent as soon as we start to think of the lasting, multi-century impacts of anthropogenic climate forcing. The fast feedbacks do not, for example, include the slow retreat of the continental ice sheets or the slow response of the Earth's surface properties and vegetation as, e.g., boreal forests slowly expand poleward. Accounting for these *slow feedbacks* leads to the possibility that the equilibrium long-term response to anthropogenic greenhouse gas emissions is larger than the IPCC projections we have focused on up until now. This more general notion of climate sensitivity is typically referred to as *Earth System senstivity*.

There is good evidence from long-term geological record of climate change that these slow feedbacks do indeed matter, and that the ultimate warming and associated changes in climate might be substantially larger than what is implied by the simple Charney definition of sensitivity implicit in the IPCC projections. For both the mid-Pliocene, roughly 2.8 MY ago, and the mid-Miocene, about 15 MY ago, global mean temperatures appear to have been warmer than would be expected from even the upper range of the estimated Charney sensitivity (4.5°C for CO_2 doubling). This suggests an earth system sensitivity that is substantially higher than the standard Charney estimate of climate sensitivity.

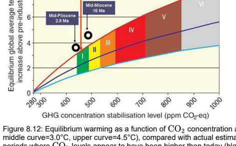
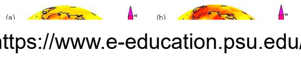


Figure 8.12: Equilibrium warming as a function of CO_2 concentration assuming a Charney sensitivity range of 3°C +/-1.5°C (lower curve=1.5°C, middle curve=3.0°C, upper curve=4.5°C), compared with actual estimates of CO_2 concentration and global mean temperature for past geological periods where CO_2 levels appear to have been higher than today (black circles).
Credit: Climate Himalaya Initiative [213]

Studies using climate models that incorporate these slow feedbacks find that the *Earth System sensitivity* is indeed substantially greater than the nominal Charney sensitivity, roughly 50% higher. Thus, a stabilization of CO_2 levels at twice pre-industrial levels over the next century might lead to a warming of 3°C over the next 1-2 centuries, but an eventual warming closer to 4.5°C once the land surface and vegetation has equilibrated to the new climate and the ice sheets have melted back to their new equilibrium configuration for the higher CO_2 concentration—a process that could take a thousand years, but perhaps substantially less.



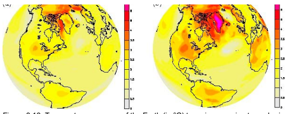


Figure 8.13: Temperature response of the Earth (in °C) to an increase in atmospheric carbon dioxide from a pre-industrial level of 280 ppm to 400 ppm. On left (a) is the projected temperature change based only on fast feedbacks, while the right (b) shows the eventual warming once slow feedbacks related to the land surface/vegetation changes fully kick in.
Credit: Institute of Physics [214]

Tipping Points

We will wrap up our discussion of climate change projections with a discussion of so-called *tipping points*. Tipping points are important because they represent possible threshold responses to forcing. While many of the climate change impacts we have looked at, like surface temperature increases, are projected to follow increasing atmospheric CO_2 concentrations in a smooth manner, there are other responses that can be more abrupt—once a certain amount of warming takes place, some component of the climate system abruptly transitions to some other regime of behavior. We saw an example of this sort of behavior in our discussion in Lesson Five (see the series of jing videos at the bottom of this Lesson Five page [215]) of the role of ice-albedo feedback in the long-term evolution of the Earth's climate system. In that case, we saw that there is the potential for more than one equilibrium state of the Earth's climate under a given state of the solar constant. When that is the case, it is possible that one or more of the steady states is unstable. In that case, a small perturbation can cause the system to spontaneously transition from its current state to a very different, other equilibrium state.

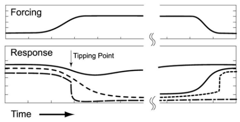


Figure 8.14: Schematic diagram of "tipping point" climate change behavior.
Credit: IPCC, 2007

We have already seen some examples of potential climate tipping points—e.g., in the potential response of the cryosphere [216] or patterns of ocean circulation [209] to ongoing warming. There are many other potential tipping points in the system, however. These include the possibility that the ENSO phenomenon might transition rather suddenly into a very different mode of behavior, or that the Indian monsoon system—whose role is so critical to fresh water availability in large parts of South Asia—might suddenly collapse. Other possibilities include one of the possible carbon cycle feedbacks alluded to previously in this lesson [217]; that a sudden release of previously frozen methane from thawing permafrost suddenly enters into the atmosphere. It is, of course, possible that other tipping points exist that we are not even aware of yet! *Can you think of any possibilities yourself?* If so—this would be another good topic to bring into this week's discussion.

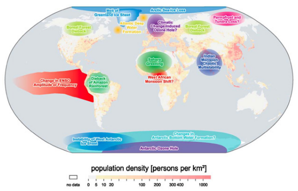


Figure 8.15: Possible Climate Change "Tipping Points".
Credit: The Copenhagen Diagnosis [190]

Lesson 8 Discussion

Activity

Please note that you will not receive a passing grade on this assignment if you wait until the last day of the discussion to make your first post.

Directions

Please participate in an online discussion of the material presented in Lessons 7 and 8: Projected Climate Change.

This discussion will take place in a threaded discussion forum in Canvas (see the Canvas Guides [18] for the specific information on how to use this tool) over approximately a week-long period of time. Since the class participants will be posting to the discussion forum at various points in time during the week, you will need to check the forum frequently in order to fully participate. You can also subscribe to the discussion and receive e-mail alerts each time there is a new post.

Please realize that a discussion is a group effort and make sure to participate early in order to give your classmates enough time to respond to your posts.

Post your comments addressing some aspect of the material that is of interest to you and respond to other postings by asking for clarification, asking a follow-up question, expanding on what has already been said, etc. For each new topic you are posting, please try to start a new discussion thread with a descriptive title, in order to make the conversation easier to follow.

Suggested topics

- What do the climate models project for the future surface temperatures? What are the two fundamental uncertainties associated with these projections? Is the warming projected to be uniform over the globe?
- What are the projected changes in precipitation and drought? What are the main causes for the large uncertainty in the precipitation projections?
- What are the projected changes in the large-scale atmospheric and ocean circulations?
- How well are the climate models able to project future changes in ENSO?
- How is the Earth's cryosphere projected to change?
- What are the projected changes of the sea level?
- What are the projected changes in tropical cyclones and extreme weather events?
- A very important point about all climate projections is that they are uncertain. One way we can attempt to assess how realistic the projections are is to compare the projections with the observations. We are in a position to do that because we have available climate projections that were done some time ago and now we have the observations that cover the projected period. Please share your thoughts on this comparison.
- What are the carbon cycle feedbacks? How many feedback mechanisms does the Earth's carbon cycle have?
- What is the difference between the Earth System Sensitivity and the Equilibrium Climate Sensitivity?
- In our current state of knowledge, what are the potential climate tipping points? Should we be concerned about these potential tipping points now or wait for a higher level of scientific certainty before we do anything?

Submitting your work

1. Go to Canvas.
2. Go to the *Home* tab.
3. Click on the *Lesson 8 discussion: Climate Change Projections*.
4. Post your comments and responses.

Grading criteria

You will be graded on the quality of your participation. See the online discussion grading rubric [19] for the specifics on how this assignment will be graded.

Lesson 8 Summary

In this lesson, we further examined potential anthropogenic climate change influences on a host of climate and meteorological phenomena. We found that:

- Arctic sea ice extent is projected to continue to decrease, and it is likely that summer sea ice will disappear if the globe warms up beyond 2°C. Observations suggest that the rate of decline is greater than what is projected by the models;
- mountain glaciers and small outlet glaciers around the world are projected to continue their decline under further warming, but there is greater uncertainty regarding the rate and magnitude of decline of the two major continental ice sheets—Greenland and Antarctic;
- while we may commit to enough warming to set in motion the ultimately disappearance of the Greenland and West Antarctic ice sheets within decades and given only modest additional warming, current generation ice sheet models suggest that the scenario could take many centuries or even millennia to play out, owing to the intrinsically slow nature of key governing processes;
- actual observations of ice mass suggest that both Greenland and West Antarctic ice sheets already have entered into a regime of negative mass balance, and it is quite possible that physical processes not well represented in current generation ice sheet models, such as ice shelf buttressing, could allow for a much faster retreat;
- there are several different components expected to contribute to global sea level rise, including the expansion of sea water with warming, the contribution of melting mountain glaciers, and ice loss from the major ice sheets; the latter component is both likely the largest and the most uncertain;
- semi-empirical models which attempt to account for all contributions to global sea level rise suggest the possibility of a meter of sea level rise or more by 2100, and perhaps as much as 5 meters by 2300;
- anthropogenic climate change is projected to lead to further increases in the incidence of specific types of extreme weather events. These include intense rainfall and snow events and flooding, and the incidence of high temperature extremes and heat waves. While the shifts in extreme weather thus far have been subtle, the projected increases are likely to be perceptible in the pattern of day-to-day-weather;
- climate change projections have traditionally neglected potentially important *carbon cycle feedbacks* which may serve to further accelerate anthropogenic climate change. For a given emissions pathway, CO_2 concentrations might increase more than indicated by the nominal CO_2 concentration scenarios because of feedbacks which tend to further increase levels of atmospheric CO_2. The impact is quite uncertain, but probably about 20-40 ppm additional CO_2 for each degree C of warming, which has non-trivial consequences for the magnitude of future warming;
- the traditional so-called "Charney" concept of equilibrium climate sensitivity may not be appropriate for assessing the full extent of anthropogenic climate change, because it neglects slow feedbacks like changes in vegetation and land surface properties, and the retreat of the continental ice sheets, which tend to further amplify the projected changes; the importance of these long-term feedbacks is suggested by the relationship between CO_2 levels and global warmth in the geological climate record. The so-called *Earth System Sensitivity*, which attempts to incorporate the amplifying effects of the slow-feedbacks, suggests a modestly (50%) greater amount of warming and associated climate change than the traditional notion of climate sensitivity;
- there are a number of phenomena, including the behavior of the cryosphere, the thermohaline circulation of the ocean, and certain carbon cycle feedbacks related to, e.g., currently frozen methane stores, which may behave in a threshold-like manner. Rather than exhibiting a steady trend in time, these phenomena may exhibit abrupt transitions in behavior in response to ongoing anthropogenic climate forcing. Other potential examples of systems that might exhibit tipping point behavior are ENSO and the Asian summer monsoon.

Reminder - Complete all of the lesson tasks!

You have finished Lesson 8. Double-check the list of requirements on the first page of this lesson [218] to make sure you have completed all of the activities listed there before beginning the next lesson.

Source URL: https://www.e-education.psu.edu/meteo469/node/111

Links
[1] http://Montclairoak.com
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