Introduction to Climate Science

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Dedication

For Ella.

Contents

Preface

This book is an introduction to climate science for undergraduate students of both science and non-science majors and for everybody interested in the topic. It is quantitative and uses many graphs and numbers, but only a minimum amount of math is required.

Climate change is a controversial topic. You can find news and opinion articles about it on a daily basis. However, not all information is trustworthy and many articles are biased by political views and ideology (Oreskes and Conway, 2010). As a consequence of this misinformation the general public is confused and its views are quite different from those of climate scientists. Whereas almost all climate scientists think that the observed global warming over the past decades has mostly or exclusively been caused by humans (Cook et al., 2013), a view that is supported by major scientific organizations such as the National Academies of Sciences and the American Geophysical Union and my personal experiences from the scientific literature and conferences, only about half of all Americans think that way (Funk and Kennedy, 2016). I have written this book to contribute to a better informed public on this topic. My goal is to make the science accessible to everybody. For this reason the textbook is free and it includes up-to-date links to trustworthy science information.

Chapter 1 discusses differences between weather and climate and provides an introduction to the interacting components of Earth's climate system. Chapters 2 and 3 summarize observations of current and past changes. Chapter 4 presents theoretical foundations and discusses relevant physical processes such as electromagnetic radiation, the greenhouse effect and Earth's energy budget. Chapters 5 and 6 provide further details in how the climate system works by covering the carbon cycle and other processes such as atmospheric and oceanic circulations and the water cycle. Chapter 7 introduces climate models and chapter 8 summarizes projections of future climate change including possible impacts on natural and human systems. Economical and ethical aspects are discussed in chapters 9 and 10, followed by chapter 11, which covers possible solutions including adaptation to and mitigation of future climate change.

You will also learn some of the language climate scientists use. Highlighted words are explained in the glossary. Simply hover with the mouse over the word and an explanatory box will emerge. Exercises and questions appear in boxes such as the one below. After reading a chapter you should be able to answer the questions. Boxes are also used for some overarching concepts such as, oxygen isotopes (Chapter 3) or the budget equation (Chapter 4). Links to lecture videos are included before the references in each chapter.

Climate science has a rich history going back 200 years to French scientist Jean Baptiste Joseph Fourier. Here, we will not delve into this history, but for those interested there is a post and documentary (1.5 h) with more information. The documentary covers not only science history but also current scientists and other aspects of climate science.

Exercises

Try it yourself and search the internet for "climate change news" or "global warming news" or similar key words and read a few (two longer or three shorter) articles or opinion pieces that sound interesting to you. You can also use an article in a paper journal or newspaper. Try to include different viewpoints and find arguments for and against human made climate change and whether or not we should do something about it.

1. Summarize each piece in one or two paragraphs and provide the title and URL (if web source) or attach a copy (if paper source).
2. Look for arguments that claim scientific facts or consensus and make a list of those. Be sure to include arguments both for and against human caused climate change.
3. Indicate which arguments you think are trustworthy and why.
4. What is your assessment of the articles? Have you learned something? Are there connections between the articles? What are the questions that remain?
5. Beyond the above pieces, what are your personal questions about climate change?

Please write in prose (except for the list of arguments) and keep your writing to 1-2 pages of text. Please bring your writing to class for discussion and be prepared to use the arguments in a mock debate. During the class discussion we split the students in two groups, one arguing for human caused climate change and that it is a bad thing, the other arguing against human caused climate change and that it is a good thing. Two students summarize the main points made by each group on the white board. The list of topics and questions of the students will be used throughout the class attempts will be made to get back and answer them based on what we'll learn.

It can be difficult to distinguish between trustworthy information and misinformation. Here are some general tips that may help you to separate the wheat from the chaff.

- **Trustworthy sources** are the Intergovernmental Panel on Climate Change (IPCC), government agencies such as the National Oceanic and Atmospheric Administration (NOAA) and the National Aero and Space Administration (NASA), and scientific organizations such as the National Academies of Sciences, Engineering and Medicine, the American Geophysical Union (AGU), and the American Meteorological Society (AMS). The IPCC reviews the state of the science with regard to climate change every 6 years or so and summarizes the current understanding in assessment reports. The latest IPCC assessment reports (AR6) were published between 2021 and 2023. The IPCC consists of three working groups: WG1: The Physical Science Basis, WG2: Impacts, Adaptation, and Vulnerability, and WG3: Mitigation of Climate Change. Each working group publishes a detailed report. The three reports are summarized in a synthesis report. These reports are extensive (1,000+ pages) and a great source for scientists but not very accessible for the general public. However, the IPCC also summarizes its findings in its Summary for Policymakers, which are shorter and more accessible. I would recommend to read those. Climate scientists from around the world contribute to these reports. This textbook makes use of the IPCC reports findings and figures.

- Articles that refer to the **peer-reviewed literature** and that include links or references that can be traced to the original work are preferable to articles that don't include references. Peer-review is an important part of modern scientific practice. As part of this process the editor of a journal selects independent reviewers that are experts in the field (peers), who then read and comment on a manuscript. These comments, which are often critical and point out weaknesses or inconsistencies, are relayed back to the authors who then can revise the manuscript taking the reviewers comments into account. Only after all reviewers agree to the publication is it that in most cases the article gets published (the editor decides). Sometimes multiple revisions are necessary until all reviewers are satisfied. Reviewers examine if the conclusions of the paper are supported by evidence presented. This process is a check that improves generally the quality of the papers and makes it more difficult to publish conclusions that are not supported by evidence. But, as all human endeavors, it is not perfect.
- Look at other articles on the same web page. If you see signs for conspiracy theories or claims that climate change is a hoax it is likely that you're on a misleading webpage.
- Compare the claims with trustworthy sources. If there are major discrepancies you have a reason to be doubtful.
- Critically examine the arguments. Are they compelling? Do the authors cherry-pick data? Do they present only one side of the argument and neglect other evidence?
- Some articles are reviewed by scientists at ClimateFeedback.org.
- If in doubt ask a scientist. Feel free to write me or another climate scientist an email.
- The following are websites that, in my opinion, misrepresent and/or distort climate science:
    - https://www.friendsofscience.org
    - http://climatesciencenews.com
    - https://www.climatedepot.com
    - https://wattsupwiththat.com
    - https://www.cato.org
    - http://www.aei.org
    - https://www.heartland.org
    - http://principia-scientific.org

Videos

All lecture videos are available here.

Lectures

Lecture slides are available here.

References

Cook J., D. Nuccitelli, S. A. Green, M. Richardson, B. Winkler, R. Painting, R. Way, P. Jacobs and A. Skuce, 2013, Quantifying the consensus on anthropogenic global warming in the scientific literature, Environmental Research Letters, 8, doi:10.1088/1748-9326/8/2/024024.

Funk C. and B. Kennedy, 2016, The Politics of Climate, Chapter 1: Public views on climate change and climate scientists, Pew Research Center.

Oreskes N. and E. Conway, 2010, Merchants of Doubt, Bloomsbury Press, New York, ISBN 978-1-59691-610-4.

Acknowledgements

The cover picture of Earth was taken by the crew of the Apollo 17 on their journey to the moon on December 7, 1972 (from nasa.gov). I'm grateful to Laurie Houston, Susanne Capalbo, Kathleen Dean Moore, and Michael Paul Nelson for their contributions to the chapters on economics and ethics. I want to thank my wife Susanne for her support and patience on many evenings and weekends that I spent writing. I appreciate student's comments, which have improved the book. I also want to thank Dianna Fisher, Gabe Higginbotham, and Devin Curtis from OSU's open textbook project for their encouragement and technical help.

1. Weather

a) Weather and Climate

Weather and climate are related but they differ in the time scales of changes and their predictability. They can be defined as follows.

Weather is the instantaneous state of the atmosphere around us. It consists of **short-term variations** over minutes to days of variables such as temperature, precipitation, humidity, air pressure, cloudiness, radiation, wind, and visibility. Due to the non-linear, chaotic nature of its governing equations, weather **predictability is limited** to days.

Climate is the statistics of weather over a longer period. It can be thought of as the **average** weather that varies slowly over periods of months, or longer. It does, however, also include other statistics such as probabilities or frequencies of ***extreme events***. Climate is **potentially predictable** if the *forcing* is known because Earth's average temperature is controlled by energy conservation. For climate, not only the state of the atmosphere is important but also that of the ocean, ice, land surface, and biosphere.

In short: 'Climate is what you expect. Weather is what you get.'

b) The Climate System

Earth's climate system consists of interacting **components** (Fig. 1). The **atmosphere**, which is the air and clouds above the surface, is about 10 km thick (more than two thirds of its mass is contained below that height). The **ocean** covers more than two thirds of Earth's surface and has an average depth of roughly 4 km. Contrast those numbers with Earth's radius which is approximately 6,400 km and you'll find that Earth's atmosphere and ocean are very thin layers compared to the size of the planet itself. In fact, they are about 1,000 times thinner. They are comparable perhaps to the outer layer of an onion or the water on a wet soccer ball. Yet all life is constrained to these thin layers. The major ocean basins are the Pacific, the Atlantic, the Indian, and the Southern Ocean. Ice and snow comprises the **cryosphere**, which includes sea ice, mountain glaciers and ice sheets on land. Sea ice is frozen sea water, up to several meters thick, floating on the ocean. Ice sheets on land, made out of compressed snow, can be several kilometers thick. The **biosphere** includes all living things on land and in the sea from the smallest microbes to trees and whales. The **lithosphere**, which is the solid Earth (upper crust and mantle), could also be considered an active part of Earth's climate sytem because it responds to ice load[1] and impacts atmospheric carbon dioxide (CO_2) concentrations and climate on long timescales through the movements of the continents.

Figure 1: The blue marble. A composite image of Earth from space. It shows all four components of Earth's climate system. The atmosphere with its complex cloud patterns. The ocean, which covers about 70% of Earth's surface. The cryosphere is visible as the white areas on the top: sea ice covering the Arctic Ocean and the Greenland ice sheet. Green colors on land and turquoise shades along the ocean's margin indicate the biosphere as forests and phytoplankton blooms. Notice in the lower left the thin layer of the atmosphere surrounding Earth. From nasa.gov.

1. The weight of the ice depresses the lithosphere, which provides a feedback to the ice by lowering its surface elevation. Traditionally it has been thought that this feedback acts on long (millennial and longer) timescales, but recent research indicates that in certain regions such as West Antarctica, where the lithosphere is thin and the mantle viscosity is low the lithosphere can respond on much shorter (decadal to centennial) timescales.

The components **interact** with each other by exchanging energy, water, momentum, and carbon thus creating a deliciously complex coupled system. Imagine water evaporating from the tropical ocean heated by the sun (Fig. 1). The air containing that water rises and cools. The water *condenses* into a cloud. The cloud is carried by winds over land where it rains. The rain sustains a forest. Trees are dark, having a low ***albedo***. This influences the amount of sunlight absorbed by the Earth. Dark surfaces absorb more sunlight and get warmer compared to bright surfaces such as desert sand or snow. Air warmed by the surface rises and affects the wind.

c) Processes

Fig. 2 illustrates some of the important processes that contribute to the complex interactions within the climate system. Earth's energy source is the sun. Both solar and terrestrial *radiation* are affected by gases, ***aerosols***, and clouds in the atmosphere. Thus, the atmospheric composition affects the heating and cooling of the earth. Heating and cooling affect the temperature and circulation of the atmosphere and oceans. Circulations of the air and sea affect temperatures and precipitation over both ocean and land, which impact the biosphere and cryosphere. Atmosphere and oceans exchange heat, water (evaporation and precipitation), and momentum. Wind blowing over the ocean pushes the surface water ahead. Air temperatures and snow fall affect the growth and melting of glaciers and ice sheets. Water from melting ice flows through rivers into the ocean affecting its salinity, its density and movement. Variations in solar irradiance can cause global climate to change. Volcanoes can eject large amounts of aerosol into the atmosphere with climatic implications. Humans are influencing the climate system through emissions of ***greenhouse gases***, ***aerosols***, and ***land use changes***.

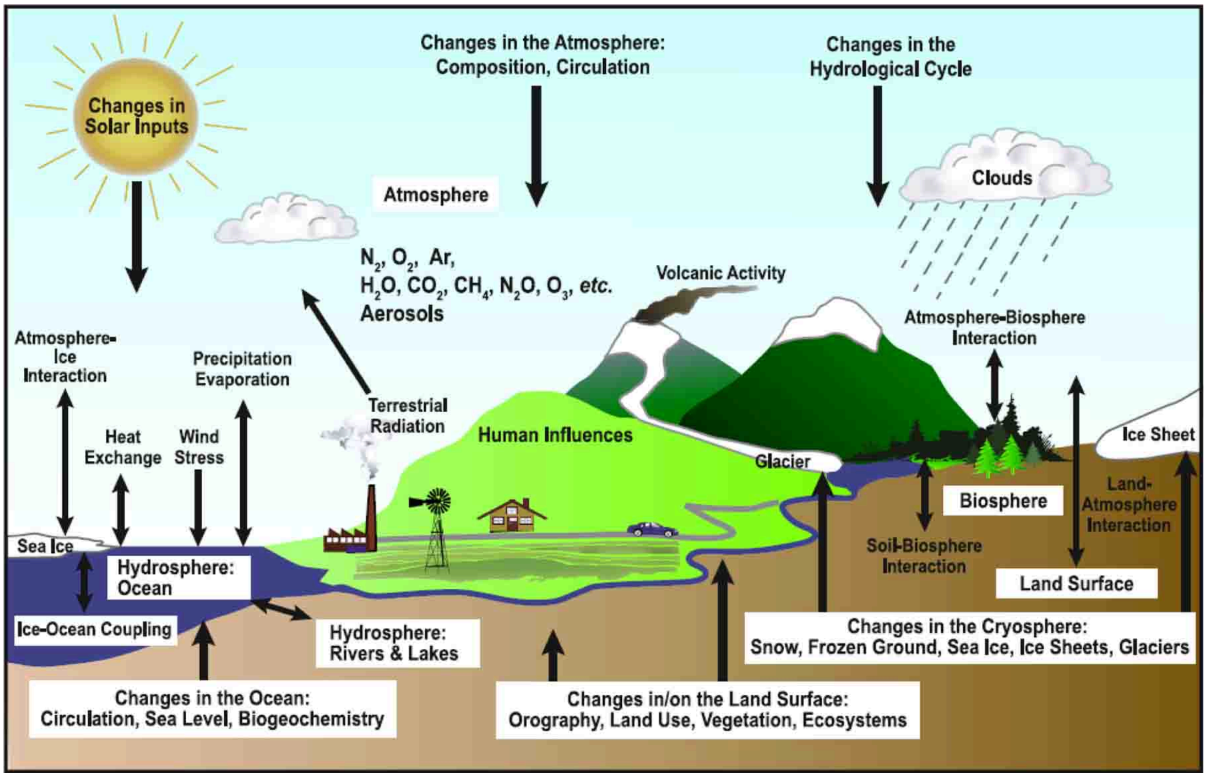


Figure 2: Schematic view of components of the climate system and processes involved in their interactions. From Le https://www.ncdc.noaa.gov/cagTreut et al. (2007).

The complexity makes studying the climate system challenging. Scientists from many disciplines contribute such as physicists, chemists, biologists, geologists, oceanographers, atmospheric scientists, paleoclimatologists, mathematicians, statisticians, and computer scientists. To me, the challenges and interdisciplinary nature of climate science are fascinating and fun. I learn something new about it every day.

Recorded Lectures

[Weather and Climate]

References

Le Treut, H., R. Somerville, U. Cubasch, Y. Ding, C. Mauritzen, A. Mokssit, T. Peterson and M.

Prather, 2007: Historical Overview of Climate Change. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Questions

- What are the two main differences between weather and climate?
- How big is Earth? How thick is the atmosphere, how deep the ocean?
- What are the components of Earth's climate system?
- List processes that cause interactions between the components: atmosphere-ocean, atmosphere-biosphere, atmosphere-cryosphere, ocean-biosphere, ocean-cryosphere.
- Name three processes (***forcings***) that can cause global climate to change.

Explore

Go to one of the following websites and explore **temporal variations** of surface temperature observations in a location of your interest. Plot today's data, yesterday's, last week's, last month's, last year's, the last decade's, as far back as you can and make some notes on what you observe. Note regular, predictable cycles such as the diurnal and annual cycle in contrast with irregular, unpredictable variations such as day-to-day and year-to-year fluctuations. What causes the regular cycles? Try to order the variations with respect to their amplitude (strength). Discuss with fellow students and instructor.

Hourly and daily data are available at: https://www.wunderground.com/history/. Enter a location of your interest. Click on the "Weekly", "Monthly", or "Custom" tabs to select different time periods. Unfortunately, when I tried it Jan. 30th, 2017 the "Custom" option didn't allow one to plot more than one year of data. However, you may be able to identify diurnal (daily) cycles in temperature. Fig. 3 shows an example. Although some nights can be warmer than some days and sometimes days and nights have the same temperature, on average days are warmer than nights. Thus the average diurnal cycle is part of climate. It is forced by the diurnal cycle of solar irradiance. Does pressure have a diurnal cycle? Answer: no. What is the typical timescale of pressure variations? Often it is about a week or so. This timescale is associated with the transition of weather systems (high and low pressure systems) passing by.

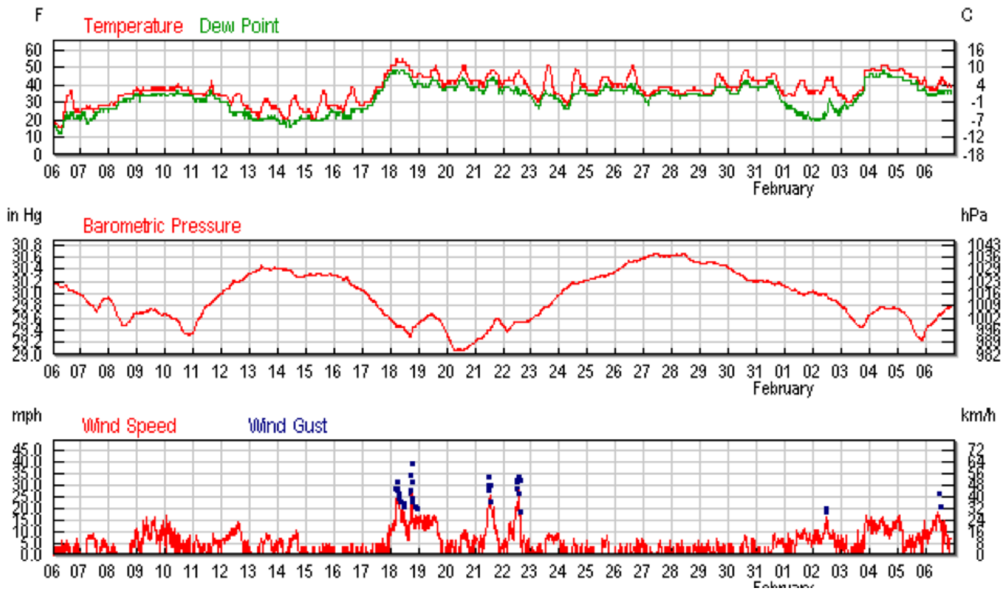


Figure 3: Weather data from Corvallis, Oregon from January 2017. In the upper panel we can see diurnal cycles in temperature. Pressure varies on longer timescales. Low pressure is associated with storms (high winds and wind gusts).

Help on finding past weather data is available at https://www.weather.gov/help-past-weather. Following the instructions and clicking on Oregon brings up http://www.wrh.noaa.gov/pqr/.

Clicking on "local", choosing the "Local Data/Records" tab, and clicking on "Daily Records for cities in Oregon" brings me to the website of the Western Regional Climate Center http://www.wrcc.dri.edu/summary/climsmor.html where I can choose a city. Now I choose "Corvallis Oregon State University" and scroll down on the left side menu to "Graph and Lister" under "Daily Data". Now select the time period (I choose 20120101 to 20160101) and click on the "Daily Precipitation" button and then "Create Graph".

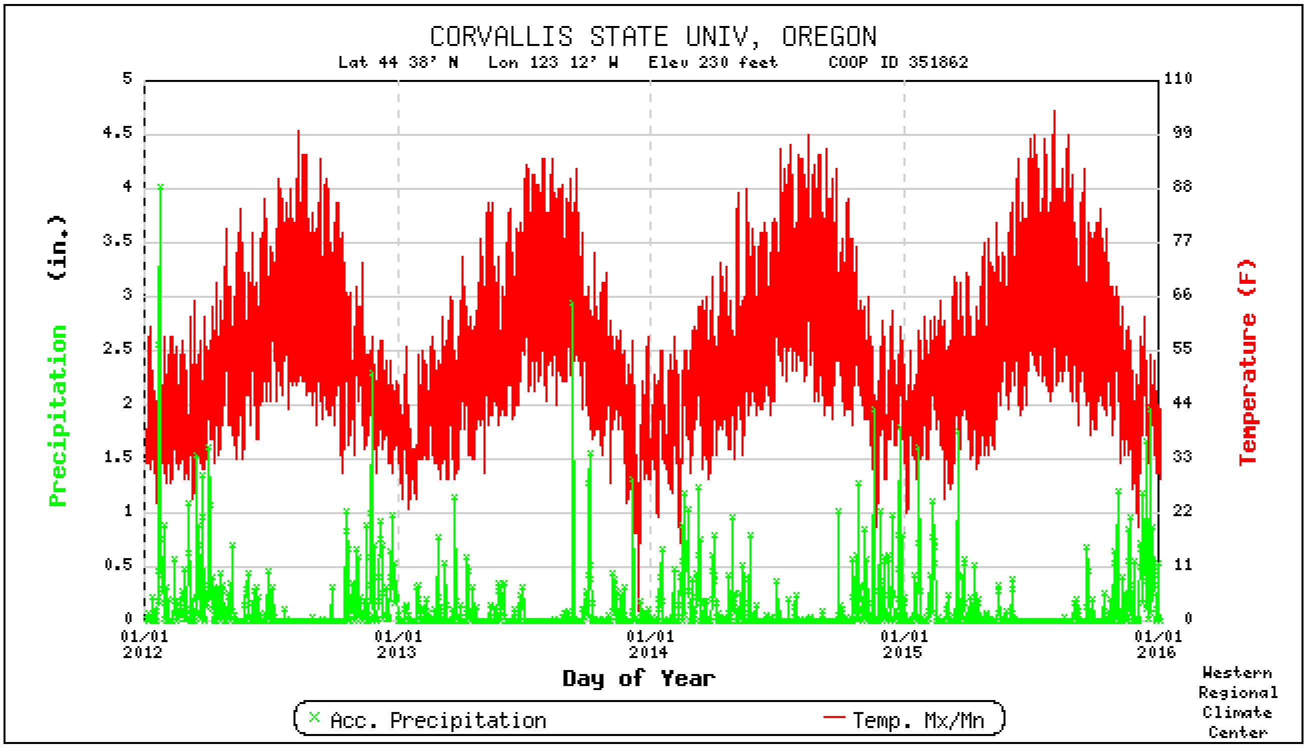


Figure 4: Local weather data (daily minimum and maximum temperature and daily precipitation) from Corvallis, Oregon, from 2012 to 2016. Large variations in temperatures from ~40°F in winter to ~65°F in summer indicate the seasonal (or annual) cycle. Precipitation, which is very episodic, also shows a large annual cycle with lots of rain in winter and less rain in summer. The mean seasonal cycle is shown in Fig. 5. From wrcc.dri.edu

Averaging many years of these weather data yields a climatology. At the WRCC website you can see plots of climatologies over 30 years by scrolling back up the left side menu and clicking on one of the "Daily Temp. & Precip." links. For Corvallis it shows average temperatures around 40°F (4°C) in winter and 65°F (18°C) in summer. It also shows a larger diurnal cycle in summer than in winter. What could be the reason for that? Answer: less cloud cover in summer. Clouds lead to cooler days and warmer nights, thus decreasing the diurnal cycle. As expected, precipitation is low in summer and high in winter. Curves are smooth in the climatology. In contrast to Fig. 4 the weather variations have been averaged out.

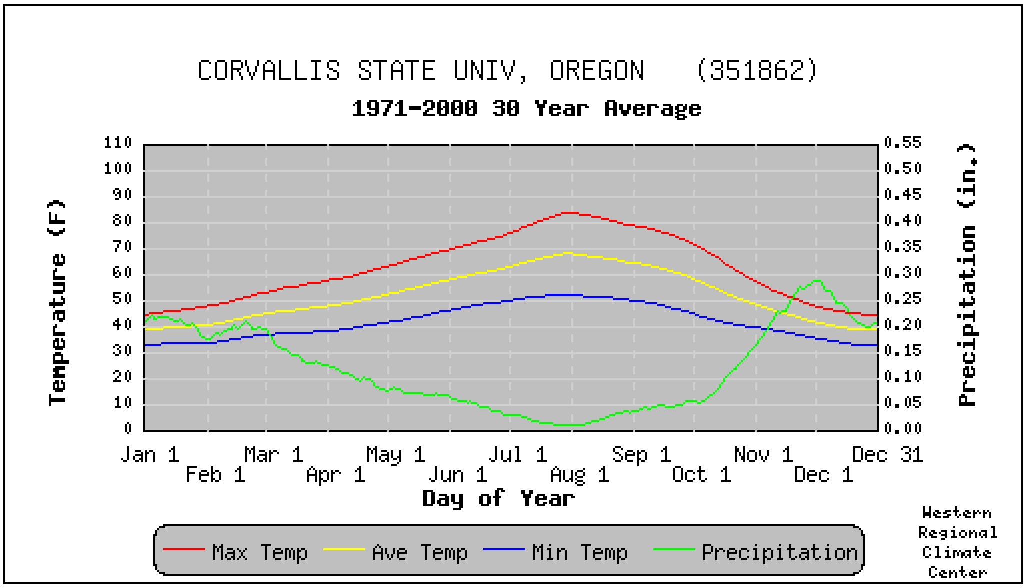


Figure 5: Climatology of temperature and precipitation in Corvallis, Oregon. The difference between the red and blue curve denotes the average diurnal cycle.

Now try a forecast of temperature for next August and next January. What is your guess? Answer: a good guess would be the climatology, i.e. 65°F for August and 40°F for January. Thus the average seasonal cycle is climate and it is predictable because it is forced by the seasonal cycle of solar irradiance. However, deviations from the climatology for a certain day next year are unpredictable weather.

Statistical properties of climate can be summarized in a frequency histogram. Let's explore those with an example. Download from the WRCC website daily average temperature data from a certain day of year, e.g. May 1st. You can do this by clicking on the "Lister" link in the left menu. Next choose start and end days (e.g. 19851001 and 20160101) to create a list of 30 years of data. Enter "wrcc35" as the password. Let's choose "Average Temperature" as the element to list. Now select a day of the year (e.g. May 1st) for both "Starting Date" and "Ending Date", and click the "Get Listing" button. This gives the following list (or vector mathematically speaking) of N=30 temperatures **T** = (T1, T2, … , TN) = (52, 52, 44, 52, 52, 54, 48, 52, 50, 56, 59, 46, 64, 58, 55, 48, 54, 54, 63, 54, 51, 54, 42, 52, 52, 48, 48, 48, 66, 56). Here N = 1 corresponds to 1985, N = 2 to 1986 and so on. In order to construct the frequency histogram we first need to choose bin sizes. I propose to choose 6 bins: 40-44, 45-49, 50-54, 55-59, 60-64, and 65-70. Now simply count how many years fall into each bin. I get 2, 6, 14, 6, 2, 1. Using python I produced the following graph (Fig. 6).

Climate change can be expressed as a change in the mean, which would correspond simply to a shift of the whole histogram to warmer or colder temperatures. However, it can also change the shape of the histogram. For example by making the distribution wider or narrower. This would increase or decrease the occurrences of extreme events. And, of course, it can both change the mean and the width. Most of the time, including in this text, discussions of climate change consider only changes in the mean. But we should keep in mind that changes to the tails of the distribution may be equally important because it is those tails that can have large impacts (heat waves, cold spells, droughts, floods, etc).

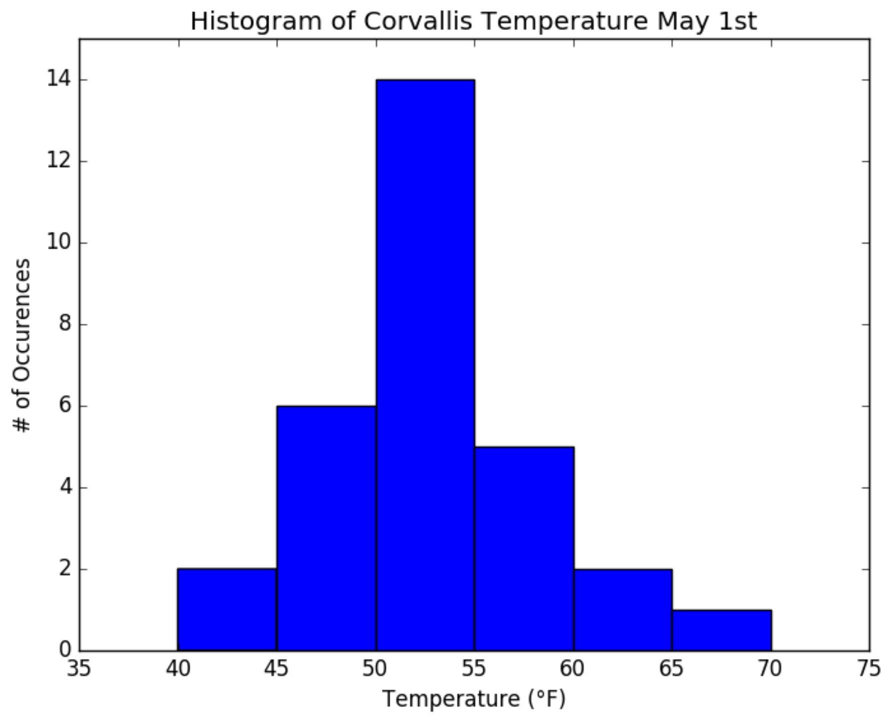


Figure 6: Histogram of 30 year (1985-2016) temperatures in Corvallis on May 1st. Most years (14 out of the 30) the temperature is between 50 and 55°F. The [mean](#) of the distribution is 52.8°F, its [standard deviation](#) of σ =5.4°F represents its width (about 2/3 of all years have temperatures within ±1 σ of the mean). The histogram, although it approximates well a Gaussian (normal) distribution, is slightly asymmetric such that very warm years (65-70) occur slightly more often than very cold extremes (35-40). The tails of the distribution are the upper and lower bins. They represent rare or extreme events. Only one year was warmer than 65°F. This year (2014) can be regarded as an extreme event. It was a record warm year in Oregon.

Now let's explore effects of **spatial scales** on climate variations. Go to https://www.ncdc.noaa.gov/cag and start with the U.S. tab selection. We want to compare the annual average temperature change in a city of your choice (e.g. Salem in Oregon), averaged over the larger region of a state, for the U.S. as a whole, and the globe (use "Globe" tab and plot "Land and Ocean", "Land", and "Ocean" separately). For "Timescale" select "Annual". Check the "Display Trend" box. This will include a trend line on the graph. Compare the trends and year-to-year variations. Are the trends larger for the global ocean or for the global land?

Compare the year-to-year variations and trends in regionally and globally averaged temperatures (Figs. 7-12) to the local diurnal cycle, day-to-day weather fluctuations (Fig. 3), and to the seasonal cycle (Figs. 4-5). Even though the global temperature trends (~0.1°C/decade) are much smaller than diurnal, day-to-day or seasonal fluctuations, which can be 10°C or more, we will see in the remainder of this course that global temperature changes of a few degrees C will have important impacts on natural systems and human societies.

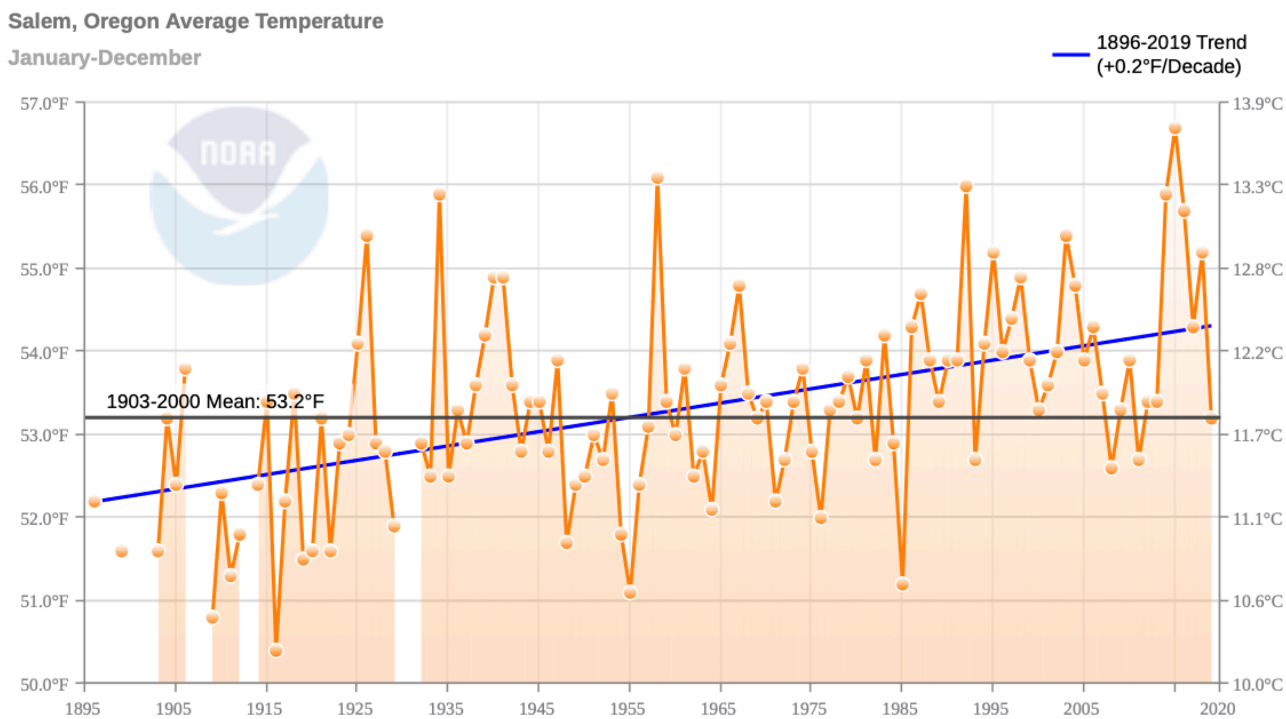


Figure 7: Annual average temperature changes in Salem, Oregon. Note the large year-to-year variations of 2-3°C and the small overall trend (blue line).

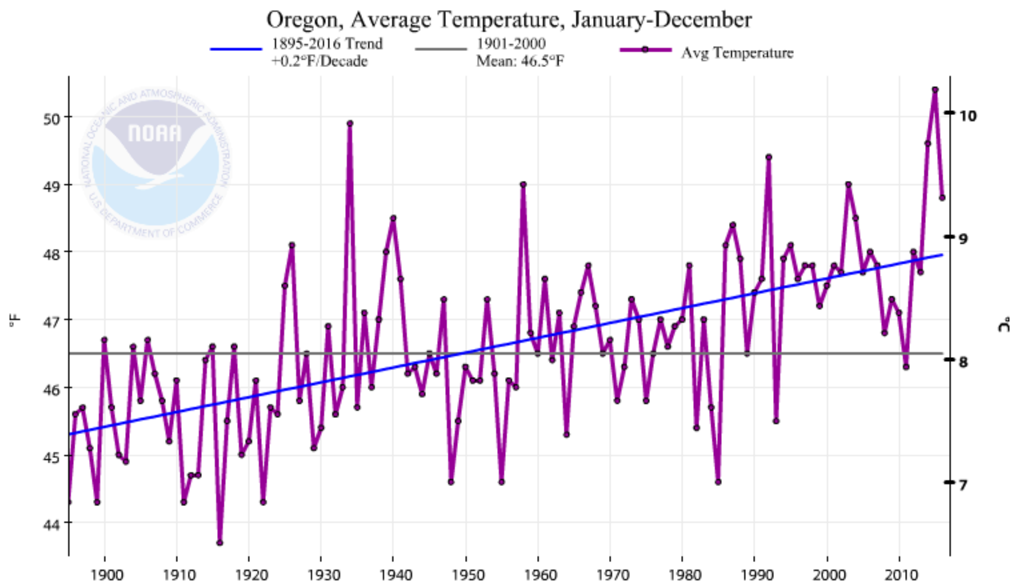


Figure 8: Averaged over the state of Oregon temperature fluctuations from year-to-year are reduced (1-2°C) compared with the figure from Salem and there is a strong (0.1°C/decade) warming trend.

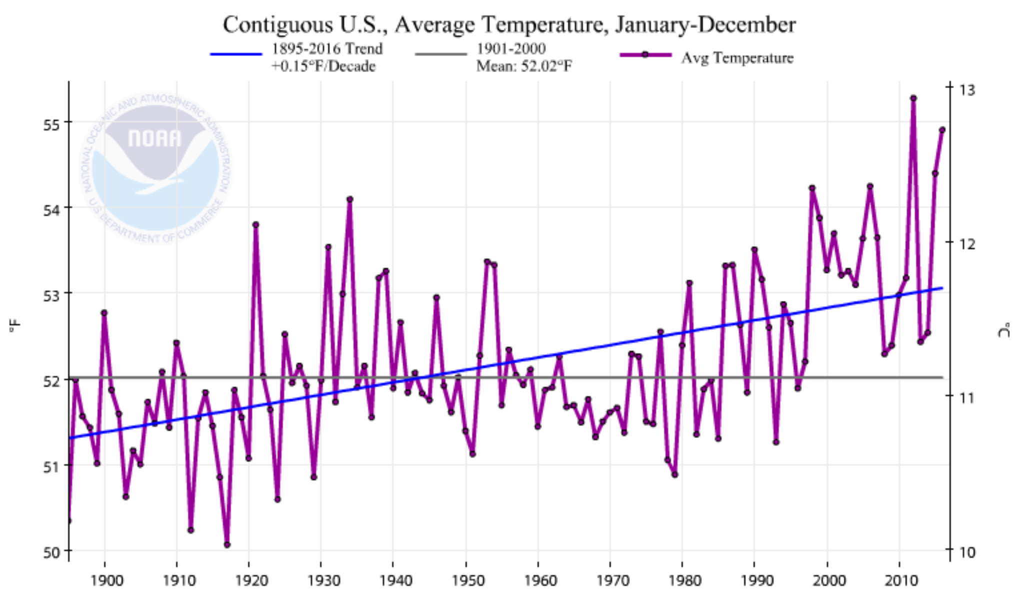


Figure 9: Averaged over the contiguous U.S. the year-to-year variability has slightly decreased compared to Oregon. The warming trend is similar but slightly lower.

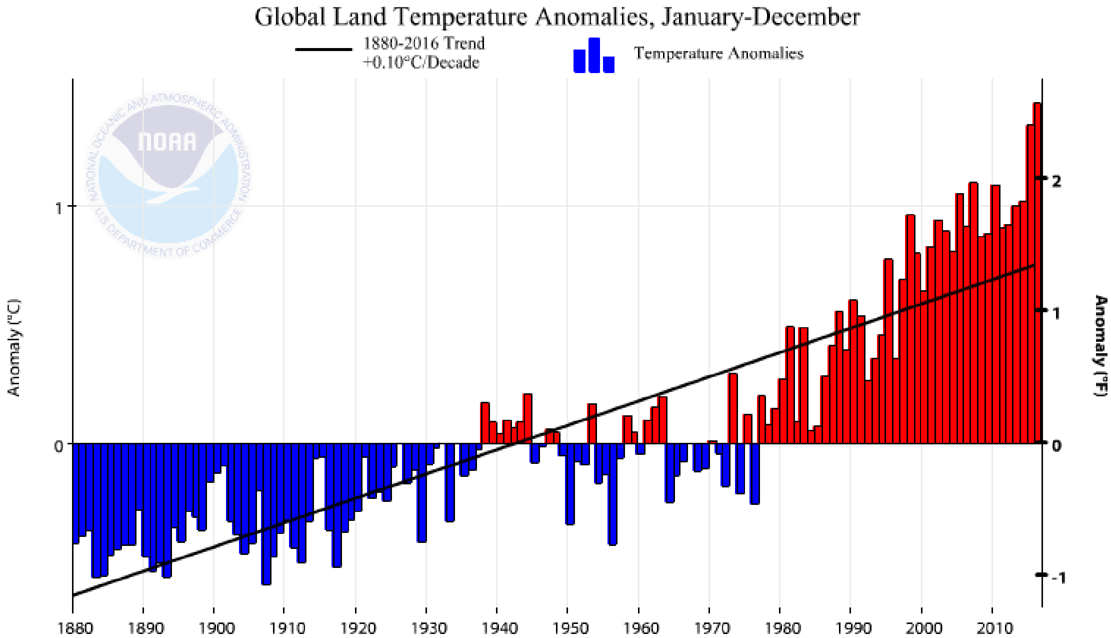


Figure 10: This figure shows anomalies (differences) with respect to the 1901-2000 average of global averaged land temperatures. Averaged over all of Earth's land areas the year-to-year variations are further reduced (~1°C) and the trend is with 0.1°C/decade slightly larger than that in the U.S.

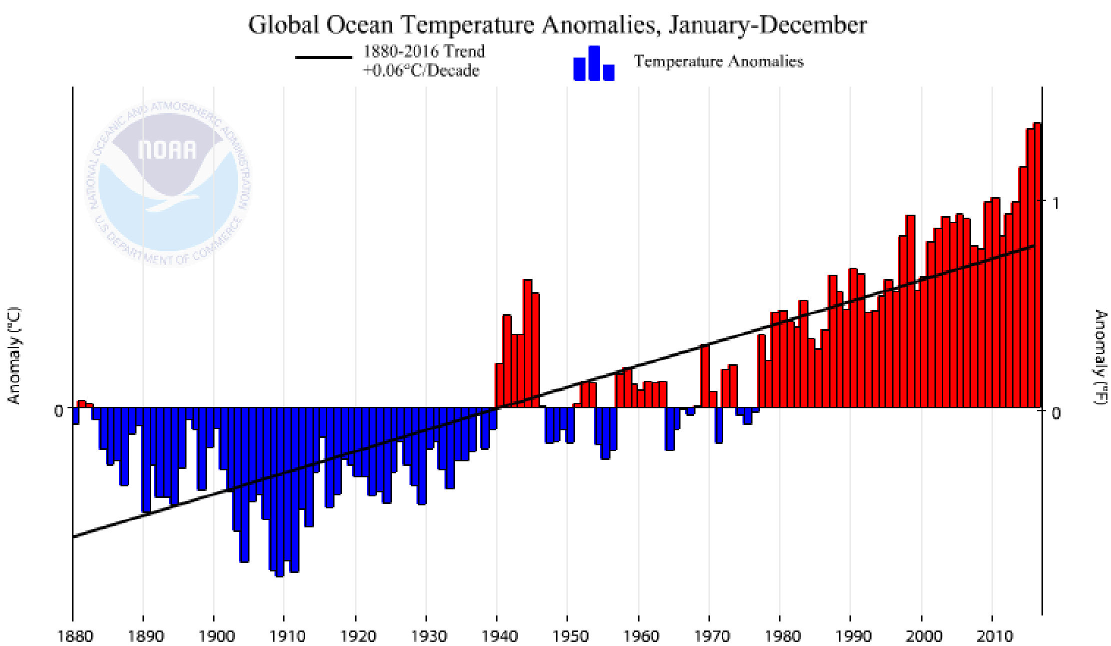


Figure 11: Averaged over the global ocean the year-to-year variations are still smaller (~0.5°C) than over land and the trend is also smaller (0.06°C/decade).

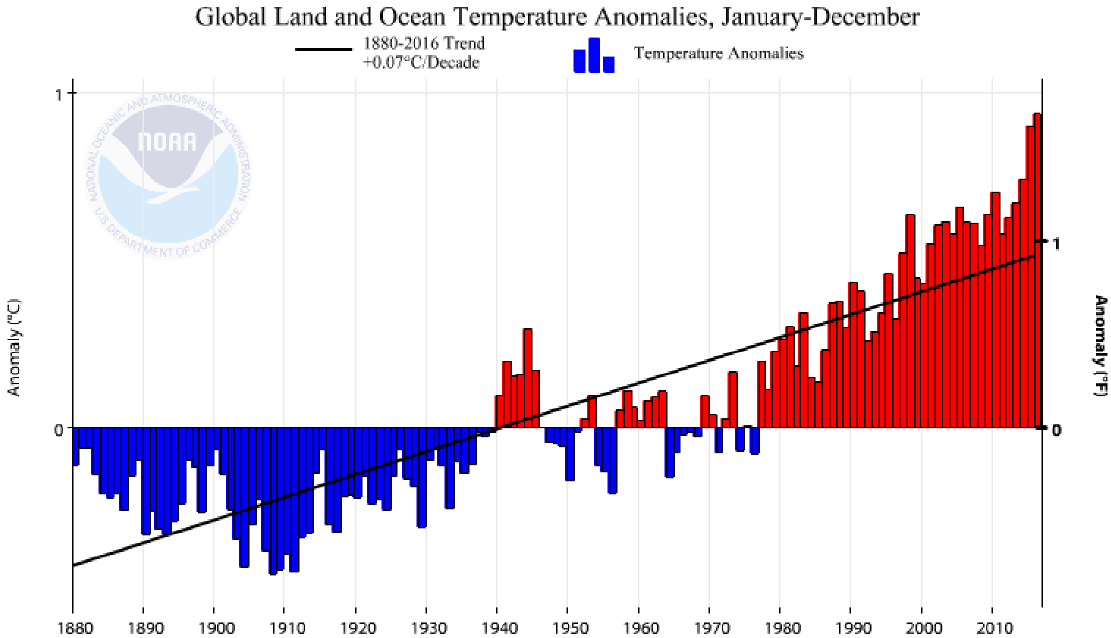


Figure 12: Averaged over land and ocean the increase in Earth's temperature over the past 100 years has been approximately 0.7°C (0.07°C/decade). This result is very robust. Several groups around the world have analyzed the available data in different ways and came to the same conclusion. This is a key figure.

2. Observations

Although Earth's climate is currently changing rapidly relative to past changes, in most regions where we live changes are slow enough that we do not notice them directly during our daily lives. However, older people may have noticed changes during their lifetimes and in some regions changes are larger and more obvious than elsewhere. In this chapter we will discuss some observations from the past 100 years and data from regions that are particularly sensitive to climate change, where the most dramatic effects have occurred. This will not be a comprehensive documentation of existing observations. Additional observations will be discussed throughout the remainder of this book.

a) Atmosphere

Observations show that climate is changing on a global scale. Surface air temperature data, averaged over the whole Earth, indicate warming of about 1°C over the last ~100 years (Fig. 1). But that warming was neither steady nor smooth. From 1880 to about 1910 there was cooling, followed by warming until about 1940. After this slight cooling or approximately constant temperatures were observed until the 1970s, followed by a rapid warming until the present. Each year's temperature is somewhat different from the next. Not all of these year-to-year changes are currently understood, but natural variations, which are not caused by humans, do play a role. For instance, strong El Niño years such as 1997-98 or 2015 show up as particularly warm years, whereas La Niña years such as 1999-2000 or 2011 show up as relatively cool. El Niño and La Niña, also known as ENSO (El Niño/Southern Oscillation), is climate variability in the tropical Pacific that impacts many other regions of the Earth.

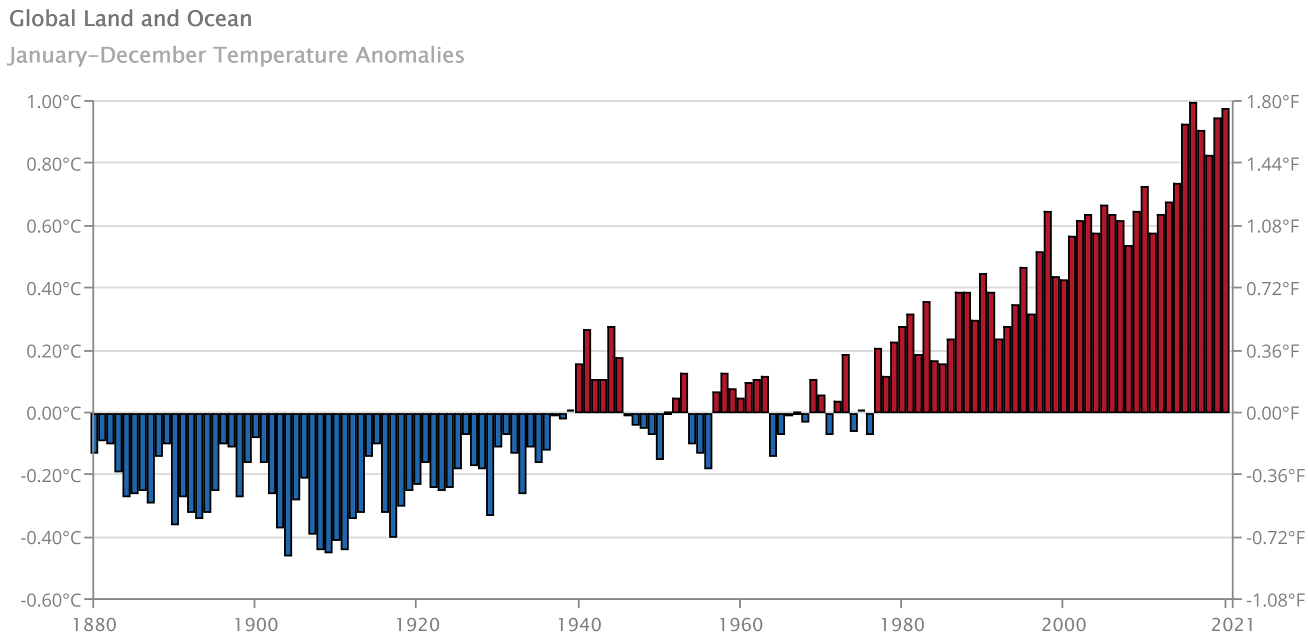


Figure 1: Global average surface temperature change from 1880 to 2018. Anomalies from the 20th century average are shown. From the National Oceanographic and Atmospheric Administration's (NOAA) web site noaa.gov. There are a few other groups around the world who calculate global average surface temperatures, e.g. the National Aeronautics and Space Administration's (NASA) Goddard Institute for Space Science (GISS), the Climate Research Unit (CRU) of the University of East Anglia, or the Berkeley Earth Surface Temperatures (BEST), and all come to very similar results. Errors due to unequal sampling have been estimated to be about 0.2°C during the early part of the record and decrease to 0.1°C during the more recent part. This is a key figure. The linear trend over the full period is 0.08°C/decade, that over the last 100 years is 0.10°C/decade and that over the last 50 years is 0.18°C/decade.

Explore Temperature Data

Go to NOAA's Climate at a Glance website to explore their global temperature data. Select "annual" to get the annual averages. Now list the 20 warmest years by clicking on the ANOMALY column in the table below until you have the warmest year listed on top.

- Which is the record warmest year?
- How many of the 20 warmest years have occurred since the year 2000?

Click on the "Display Trend" button and enter different start and end dates.

- What is the trend over the last 50 (last 100) years?

Interested in how temperature measurements from long ago have been recorded? Have a look at this fascinating interactive article.

The increase in atmospheric temperatures over the last 100 years has not been uniform everywhere. Fig. 2 shows that temperatures over land changed more than over the oceans and the Arctic warmed more than the tropics. These patterns are called ***land-sea contrast*** and ***polar amplification*** and are quite well understood and simulated in climate models as we will see later. The warming is indeed almost global. The only exception is the northern North Atlantic, which has been slightly cooling for reasons we will discuss later.

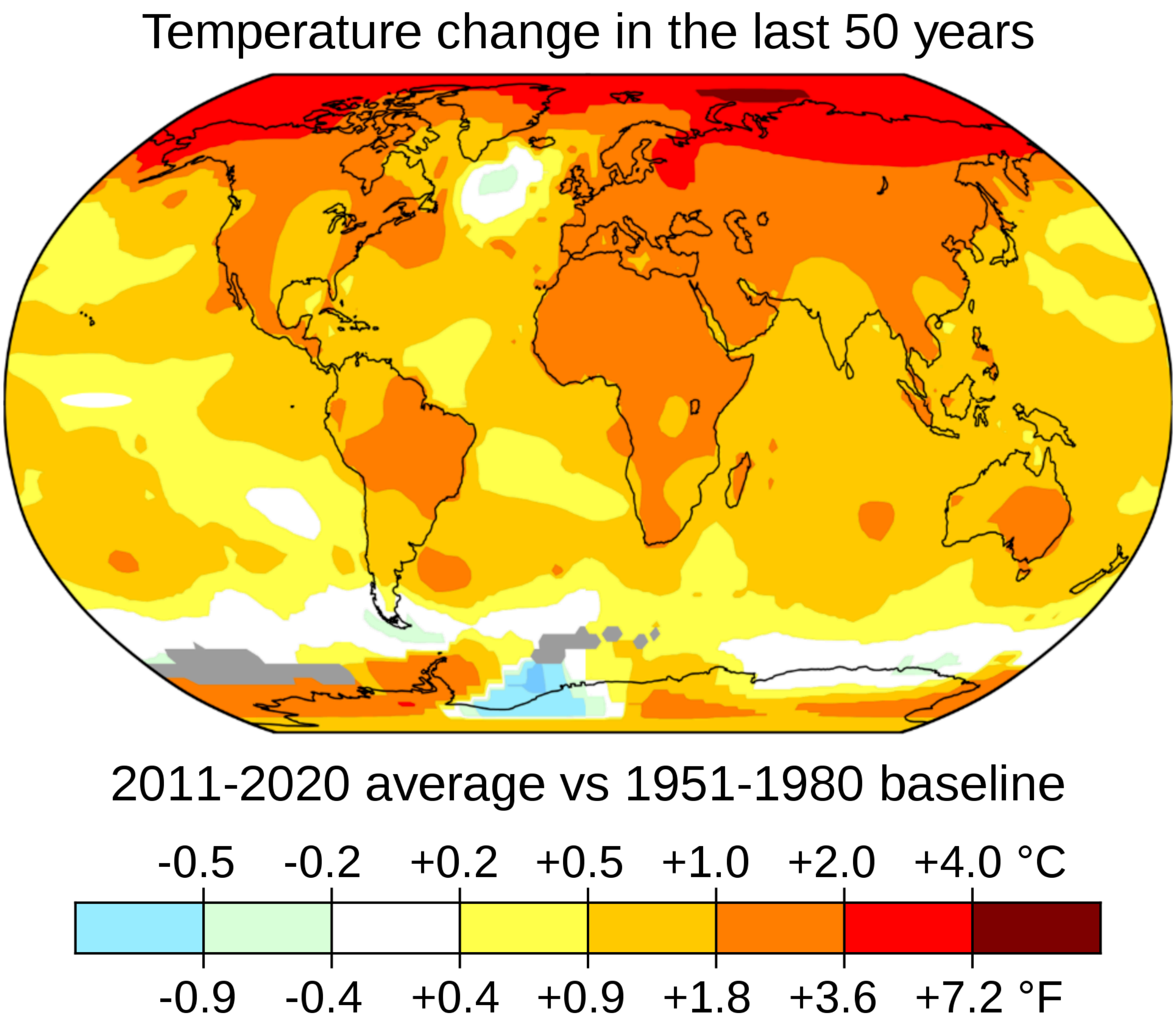


Figure 2: Global Map of Recent Surface Temperature Change. From Wikipedia's [Climate Change] based on data from NASA's Godard Institute for Space Studies (GISS).

Analysis of thousands of temperature records into an estimate of global mean temperature change such as Fig. 1 is not trivial. Issues such as changes in station locations, instrumentation and data coverage have to be taken into account. The fact that five different groups analyzing the data using different methods come to the same conclusions suggests that the results are robust. The reliability of the data has been demonstrated and it has been shown that station locations (e.g. urban versus rural) don't matter. See [this blog](#) for more discussion.

The global warming trend has made the probability of warm and extreme warm temperatures more likely and the probability of cold and extreme cold temperatures less likely. E.g. extremely warm (more than 3°C warmer than the 1951 to 1980 average) summer temperatures over Northern Hemisphere land areas had only a 0.1% chance of happening from 1951 to 1980, whereas during the decade from 2005 to 2015 the chance of those extreme warm temperatures to occur was 14.5%. Conversely, relatively cold summer temperatures (less than 0.5°C cooler than the 1951-1980 average) that happened about every third year from 1951 to 1980 only occurred 5% of the time during 2005 to 2015. See [this article](#) and [this blog](#) for additional information and graphs.

b) Cryosphere

One of the most sensitive regions to climate change is the **Arctic**. **Sea ice** cover there has decreased dramatically over the past 40 years particularly in late summer (Fig. 3). Arctic sea ice experiences a strong seasonal cycle. In late winter it covers about $15.4{\times}10^6$ km^2 and decreases to about $6.4{\times}10^6$ km^2 in late summer (Fig. 4). Therefore, the relative changes are larger in late summer, with a reduction of about 46 % or $2.9{\times}10^6$ km^2 from 1980 to 2015. In winter the reduction was only about 9 % or $1.5{\times}10^6$ km^2. The lost area of summer Arctic sea ice is more than 10 times the size of Oregon (255,000 km^2) or four times the size of France (~650,000 km^2).

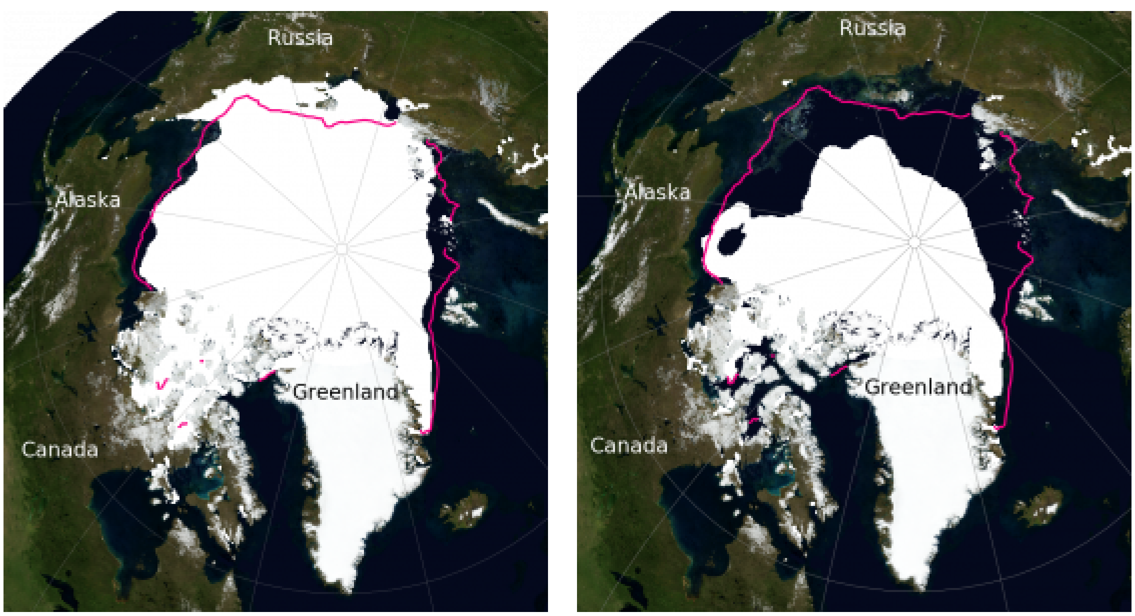
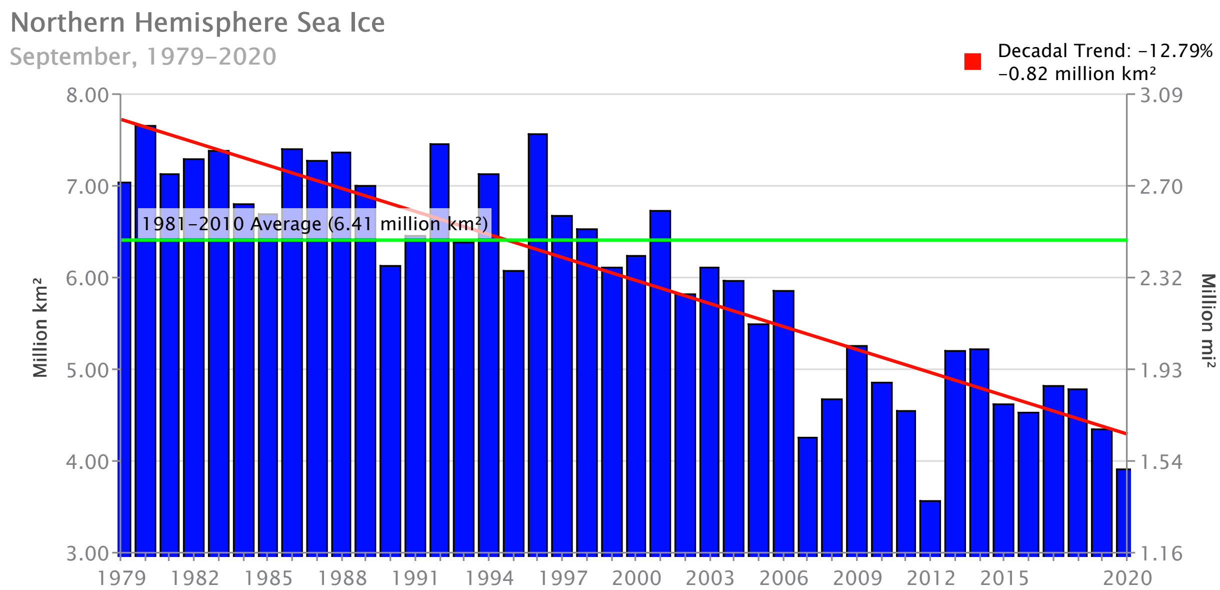


Figure 3: Arctic sea ice extent in September 1979 (left) and 2020 (right) from satellite observations. The purple line denotes the median ice extent from 1981-2010. In 1979 the ice extent

was 7.1 million sq km, in 2020 it was 3.9 million sq km. Images from the National Snow and Ice Data Center (NSIDC).

In the **Antarctic,** sea ice experiences an even larger seasonal cycle but it has changed less over the last 40 years compared to the Arctic. Southern hemisphere sea ice has slightly increased. Note that year-to-year fluctuations are larger Antarctic than in the Arctic, whereas the long term trends in the Antarctic are much smaller than in the Arctic. This combination of larger short-term fluctuations and a smaller long-term trend makes Antarctic sea ice trends less statistically significant. A trend is statistically significant if it is larger than the uncertainties from short term fluctuations.


Northern Hemisphere Sea Ice
September, 1979–2020
Decadal Trend: −12.79%
−0.82 million km²
1981–2010 Average (6.41 million km²)
Million km²
Million mi²
8.00
7.00
6.00
5.00
4.00
3.00
3.09
2.70
2.32
1.93
1.54
1.16
1979
1982
1985
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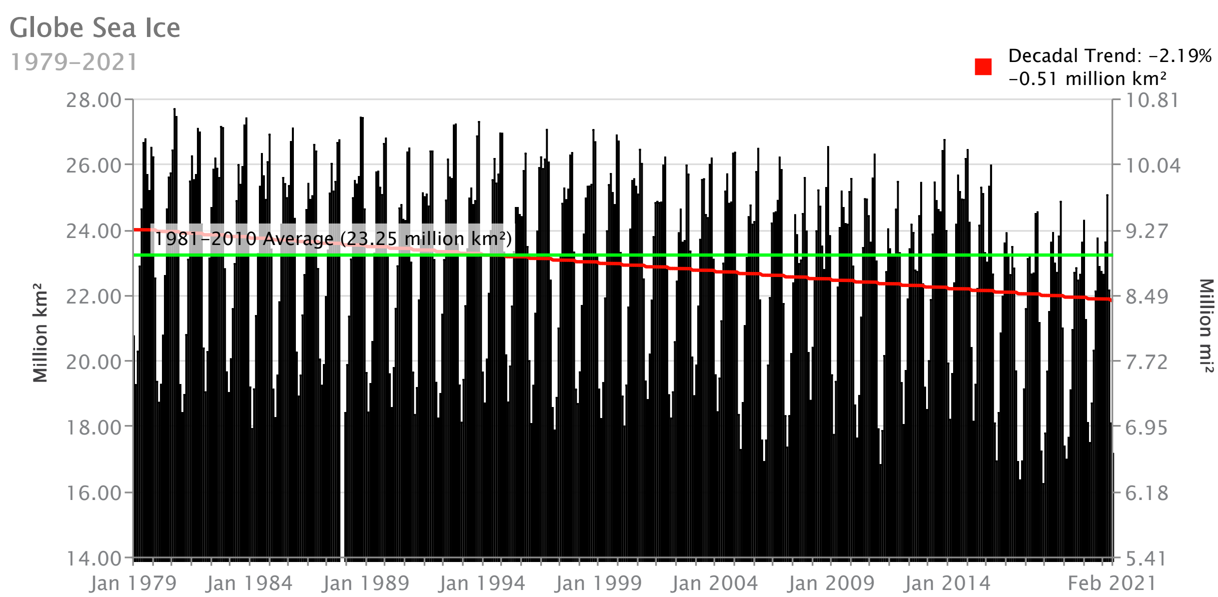
2006
2009
2012
2015
2020


Globe Sea Ice
1979–2021
Decadal Trend: −2.19%
−0.51 million km²
1981–2010 Average (23.25 million km²)
Million km²
Million mi²
28.00
26.00
24.00
22.00
20.00
18.00
16.00
14.00
10.81
10.04
9.27
8.49
7.72
6.95
6.18
5.41
Jan 1979
Jan 1984
Jan 1989
Jan 1994
Jan 1999
Jan 2004
Jan 2009
Jan 2014
Feb 2021

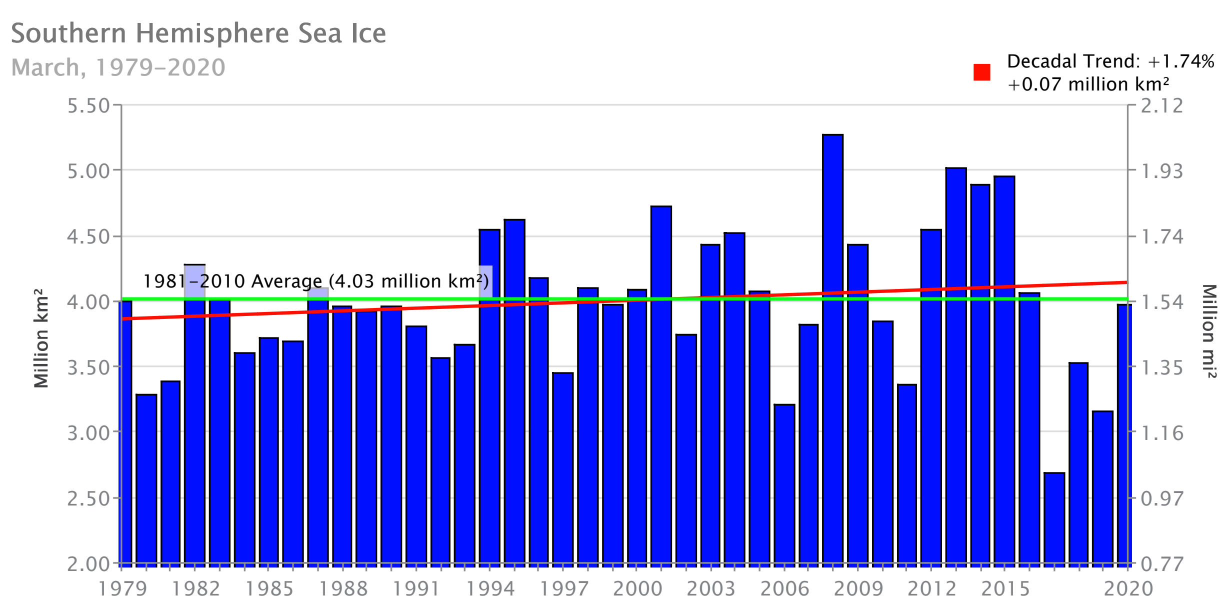

Southern Hemisphere Sea Ice
March, 1979–2020
Decadal Trend: +1.74%
+0.07 million km²
1981–2010 Average (4.03 million km²)
Million km²
Million mi²
5.50
5.00
4.50
4.00
3.50
3.00
2.50
2.00
2.12
1.93
1.74
1.54
1.35
1.16
0.97
0.77
1979
1982
1985
1988
1991
1994
1997
2000
2003
2006
2009
2012
2015
2020

Figure 4: Changes in late summer sea ice extent in the Arctic (top) and Antarctic (bottom), and global monthly sea ice extent (center). From NOAA.

Globally, Earth has lost about 2 million square kilometers of sea ice from 1980 to 2015, which is about 10% of the total. Compare that area to that of your favorite state or country.

Explore Sea Ice Changes

Go to the National Snow and Ice Data Center's (NSIDC) website and click on a few years to see how Arctic sea ice cover has changed over the year. To get a time series for the current month go to their sea ice index site and click on the Monthly Sea Ice Extent Anomaly Graph in the lower right corner.

- By how much has the sea ice cover decreased since the 1980s? Estimate the decrease both in relative terms (percentage) and in absolute terms (million square kilometers).

An animation is available here.

Mountain glaciers are also sensitive to climate change. Fig. 5 shows an example from Muir Glacier, which has retreated dramatically since 1941. This is typical for most glaciers around the world. In fact, only a small number of glaciers show advances, whereas the vast majority of glaciers melt and retreat from the valleys up into higher elevations.

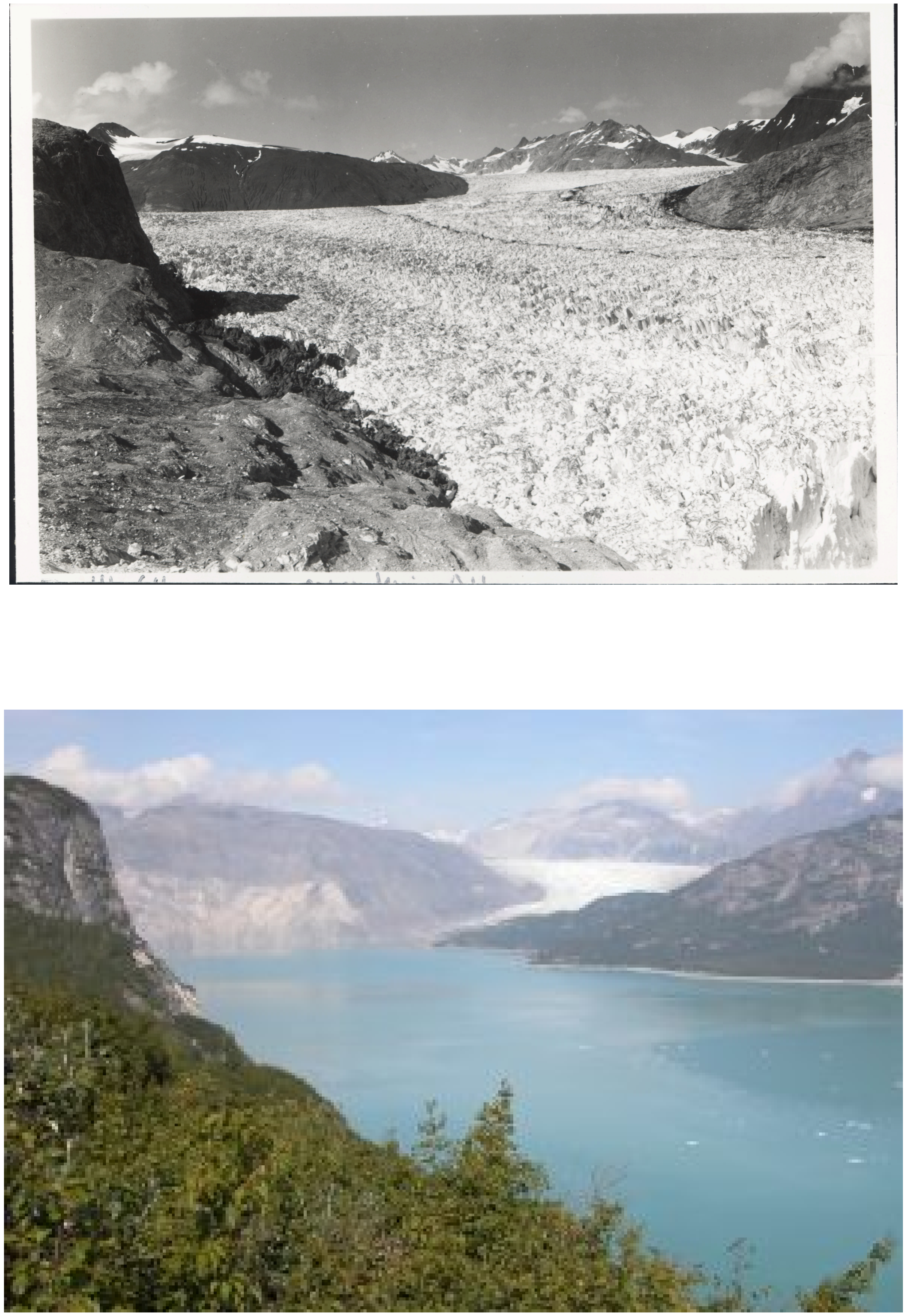


Figure 5: Muir Glacier in Glacier Bay National Park and Preserve, Alaska. The picture on top is from 1941, that on the bottom from 2004. From the National Snow and Ice Data Center (NSIDC).

The World Glacier Monitoring Service (WGMS) has compiled information on hundreds of glaciers world-wide. Fig. 6 shows that since 1980 glaciers in all regions have been losing mass with an acceleration of loss in recent years. Watch this video of glacier changes in Iceland.

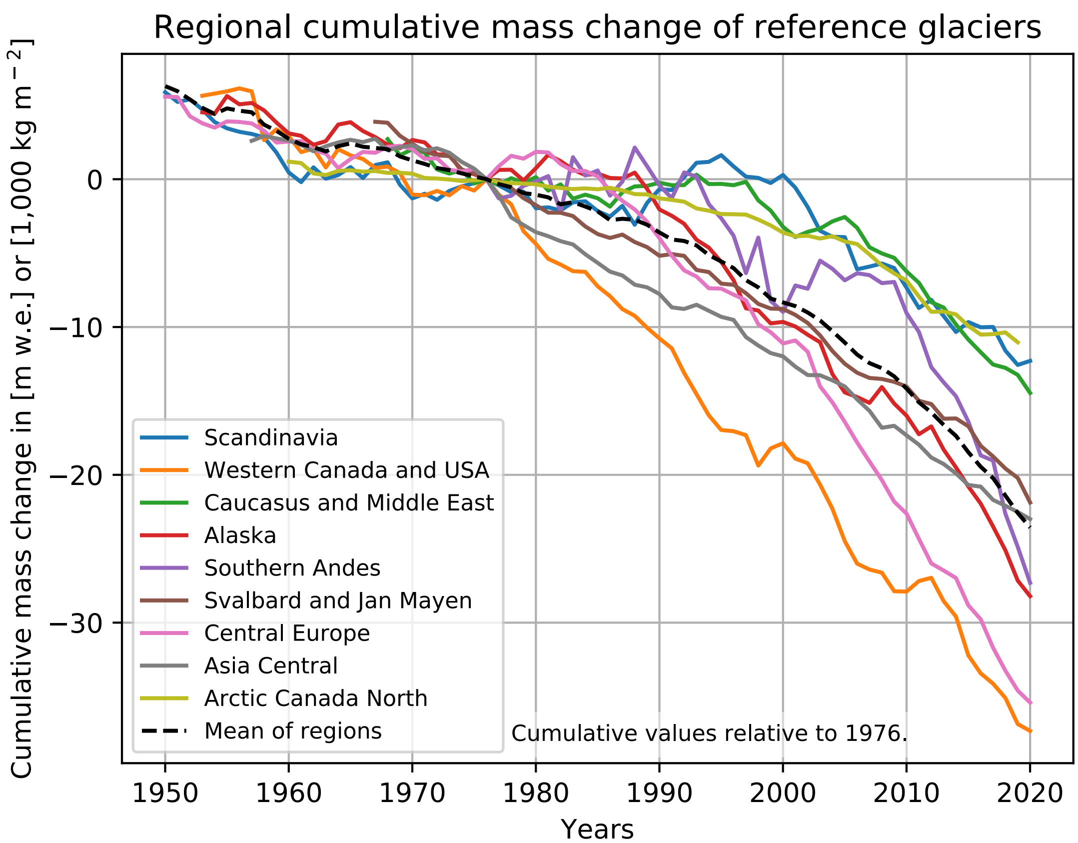


Figure 6: Cumulative mass balance in meters water equivalent (w.e.) of mountain glaciers. From the World Glacier Monitoring Service.

Explore Glacier and Ice Sheet Change

Go to the Glacier Browser and select a glacier of your choice.

- What do you observe?

Explore Greenland ice sheet change with this interactive chart, which shows the surface area experiencing melting. Click on a few years in the early part of the record (e.g. 1980s) and in the more recent part (e.g. 2010s).

- How has the melt area changed?

Ice sheets are also melting. Observations from the Gravity Recovery and Climate Experiment (GRACE) satellites, which measure very precisely Earth's gravity field and can detect changes in mass, show that since 2002 the Greenland ice sheet has lost about 4,000 Gt of mass, and the Antarctic ice sheet has lost about 2,500 Gt (Fig. 7). Here is a presentation about Greenland melting. Melting of the Greenland ice sheet contributes currently about 0.8 mm/yr to global sea level rise. Antarctica's contribution is 0.4 mm/yr and mountain glaciers add about 0.6 mm/yr, for a total of 1.8 mm/yr sea level rise from melting of ice (IPCC 2019).

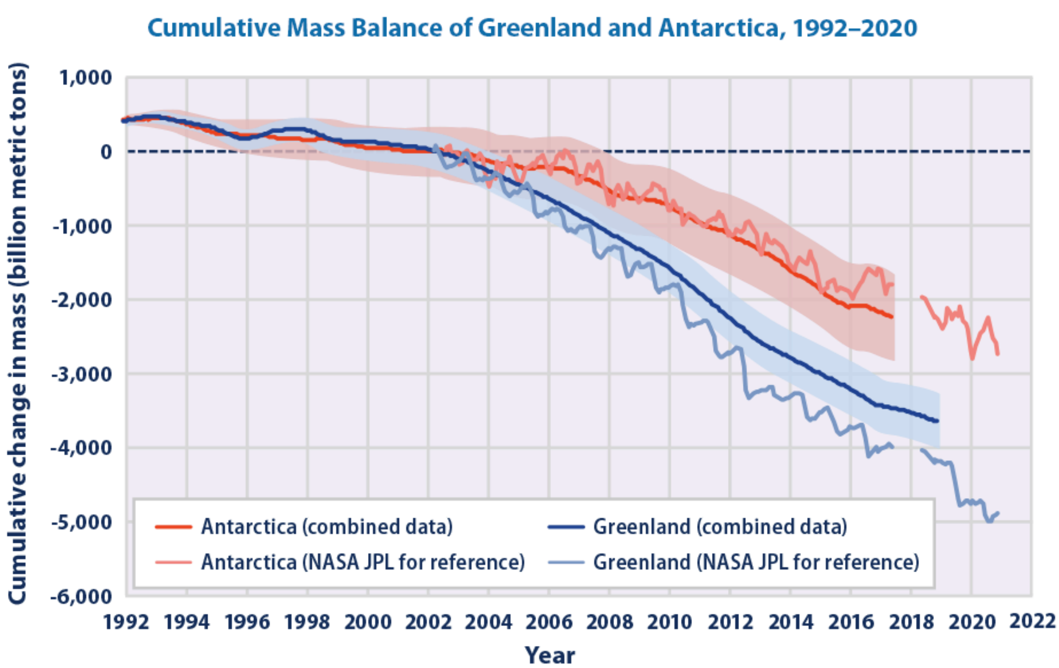


Figure 7: Changes in the Antarctic (top) and Greenland (bottom) ice sheet mass. Pale lines are from satellite gravity measurements. From the US Environmental Protection Agency (EPA; epa.gov). Satellite data are from NASA.

Box 1: Rates of Change

The rate of change of a variable X between two points in time t_1 and t_2 can be calculated as the difference (denoted by the greek letter delta Δ) of the value of the variable at time t_2 minus the value of the variable at time t_1 $\Delta X = X_2 - X_1$ divided by the difference in time $\Delta t = t_2 - t_1$

$$\frac{\Delta X}{\Delta t} = \frac{X_2 - X_1}{t_2 - t_1} \ . \quad \text{(B1.1)}$$

Thus the units of the rate of change are the units of the variable divided by time. You can determine the rate of change from a timeseries graph such as Fig. 1 or Fig. 6 by selecting two points in time on the horizontal axis and reading the corresponding values of X_1 and X_2 from the vertical axis. Using Fig. 1 as an example our variable will be the temperature anomaly T. Choosing t_1 = 1940 and t_2 = 2010 we can read off T_1 = 0°C and T_2 = 0.7°C. Thus, ΔT = 0.7°C, Δt = 70 years and the rate of change ΔT /Δt = 0.01°C/yr = 0.1°C/decade.

Obviously, calculating the rate of change in this way the resulting value will depend on the two times picked. The rate of change of a whole set of data points can be calculated by assuming a linear relationship and minimizing the distance of all points from a straight line $X = S \times t + I$, where S is the slope and I the intercept with the vertical axis. This is called linear regression. Simple formulae to calculate S and I can be found here. Linear regressions are commonly used to estimate rates of change. E.g. the straight lines in Fig. 4 in this chapter and Figs. 7-12 in chapter 1 have been calculated using the formulae for linear regression. Regression lines are also often referred to as trend lines.

c) Ocean

Subsurface temperature measurements in the oceans document warming over the last 60 years (Fig. 8). The ocean's **heat content** has increased by about 30×10^{22} Joules during that time. The heat content of the ocean is its temperature T times the heat capacity of water c_p = 4.2 J/(gK). Prior to 2005 subsurface temperature measurements were more limited in space and time because they were taken from ships by lowering CTD (conductivity, temperature, depth) instruments on a cable into the ocean. Since 2005 autonomous, free-drifting Argo floats measure temperature, salinity, pressure, and velocity of the upper 2 km of the water column. Currently there are about 4,000 floats out there, which provide much better spatial and temporal coverage than the previous ship-based measurements.

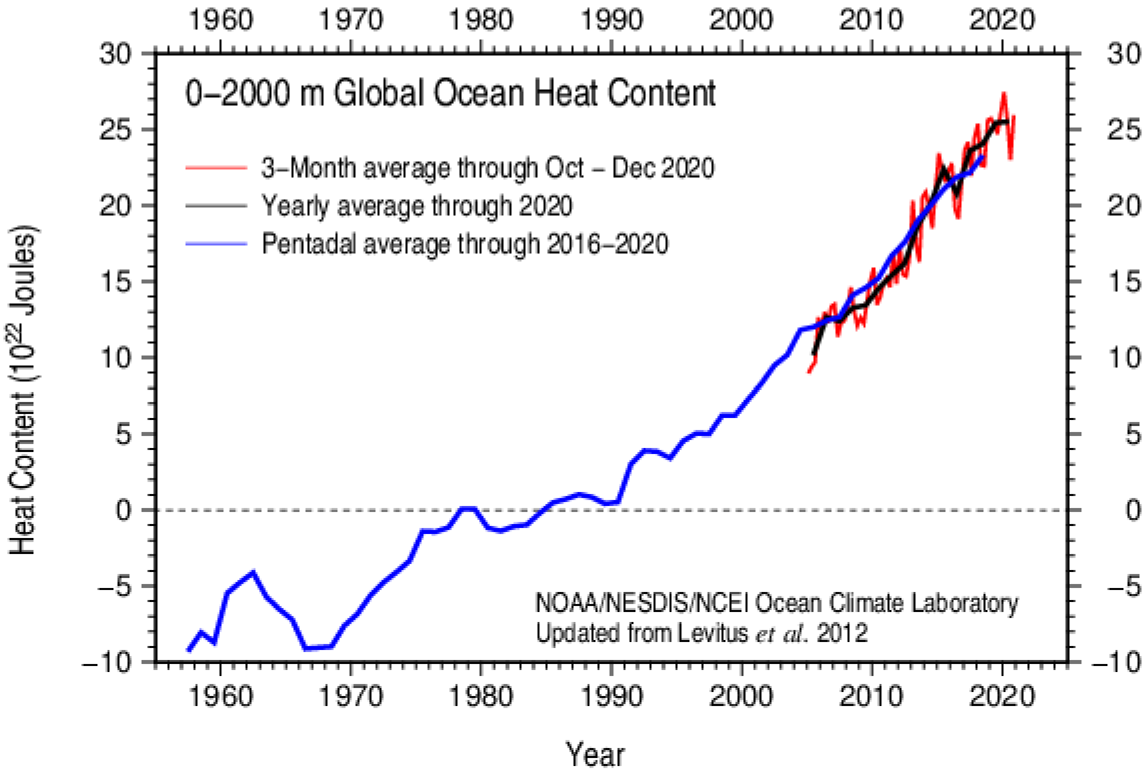


Figure 8: Observational estimates of global ocean heat content (0-2 km) from NOAA. From 1960 to 2005 ocean temperatures were measured mainly using research ships (blue line). Since 2005 autonomous ARGO floats have increased the data density both in space and time (red and black lines).

The melting of mountain glaciers and ice sheets leads to increased runoff into the ocean, which contributes to **sea level** rise (Fig. 9). Sea level rise is also caused by warming sea water, which causes expansion and by increased runoff from pumping of groundwater out of aquifers. Estimates based on tide gauge records indicate that sea level has risen by about 20 cm from the 1870s to the year 2000 and another 6 cm since. The melting of mountain glaciers and ice sheets contributes about similarly to the current sea level rise, but if current trends continue it is likely that many mountain glaciers will completely disappear and the large ice sheets will contribute more and more to global sea level rise. Note that sea level rise is not spatially uniform.

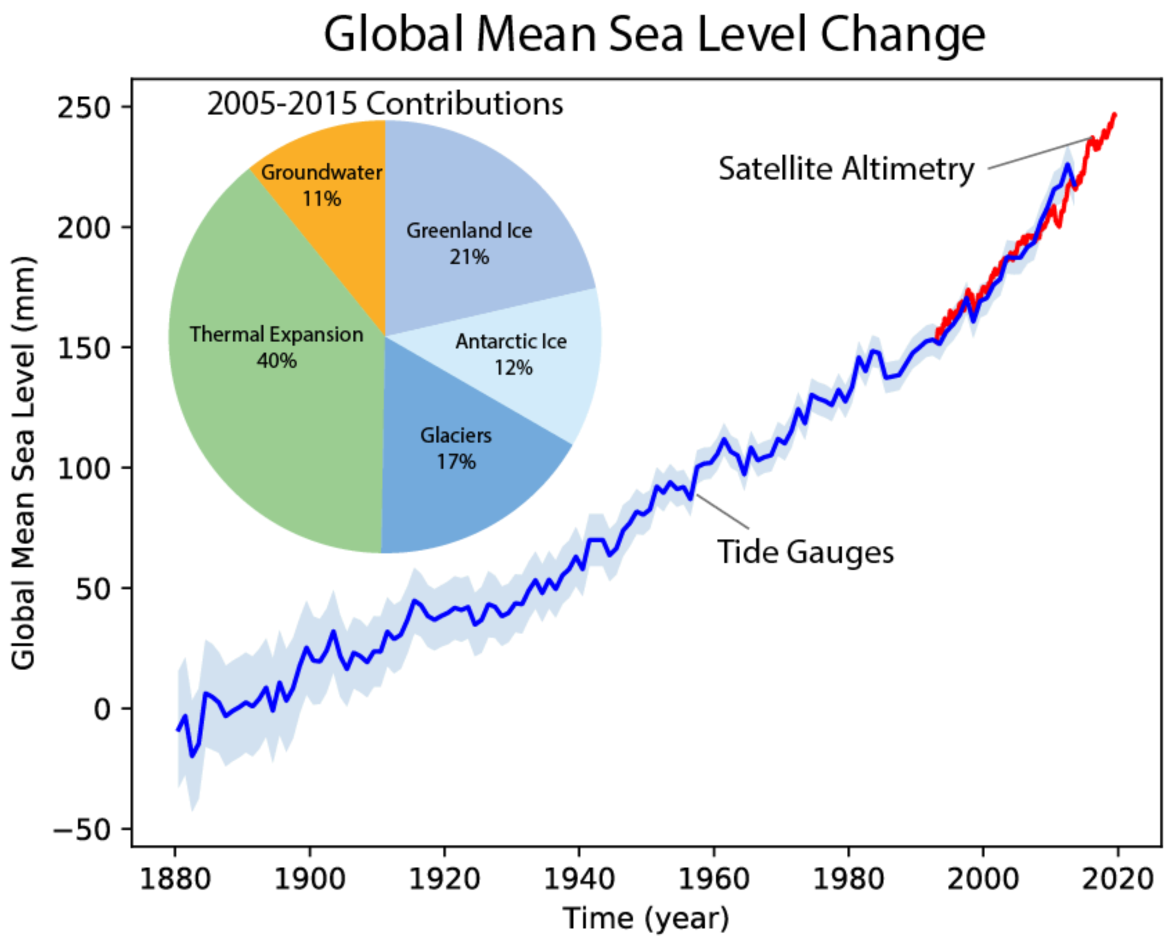


Figure 9: Observed global mean sea level estimated from satellites (red) and tide gauges (blue). Data from nasa.gov and csiro.au. The pie chart shows the relative contributions from melting of land ice (1.8±0.1 mm/yr), thermal expansion of sea water (1.4±0.3 mm/yr) and groundwater pumping (0.4 mm/yr; estimated as the residual) during 2005-2015 (IPCC, 2019). The total rate of sea level rise during this period (3.6±0.5 mm/yr) is larger than during earlier periods. This acceleration of sea level rise is caused by recently increased melting of the Greenland and Antarctic ice sheets.

Explore Sea Level Changes

Goto noaa.gov and explore local sea level changes from tide gauges.

- Where is sea level increasing?
- Where is it decreasing?
- Select a tide gauge. What is the time period covered?

Here is a map of sea level measured from satellites.

- What time period is covered by the satellite data?
- Where is sea level increasing?
- Where is it decreasing?

d) Biosphere and Carbon Cycle

Plant and animal species have been observed to move poleward and upward in the mountains (Parmesan and Yohe (2003). This response is consistent with global warming and the tendency of organisms to stay in a temperature range to which they have adapted. Flowering dates of many plants have also shifted earlier in the spring such as for cherry blossoms.

Carbon dioxide (CO_2) has been measured in the atmosphere since 1958 at Mauna Loa Observatory in Hawaii (Fig. 10). At that time concentrations were just below 320 parts per million (ppm). Subsequently they increased to values of just over 400 ppm today. That is a 25 % increase. Overlaid on the long term trend is a seasonal cycle. Growth of the terrestrial biosphere in northern hemisphere spring leads to CO_2 drawdown and decay of organic matter such as fallen leaves increases CO_2 in the fall. As we will see later, CO_2 is an important greenhouse gas, and its increase over the past decades is the main cause of the recent global warming.

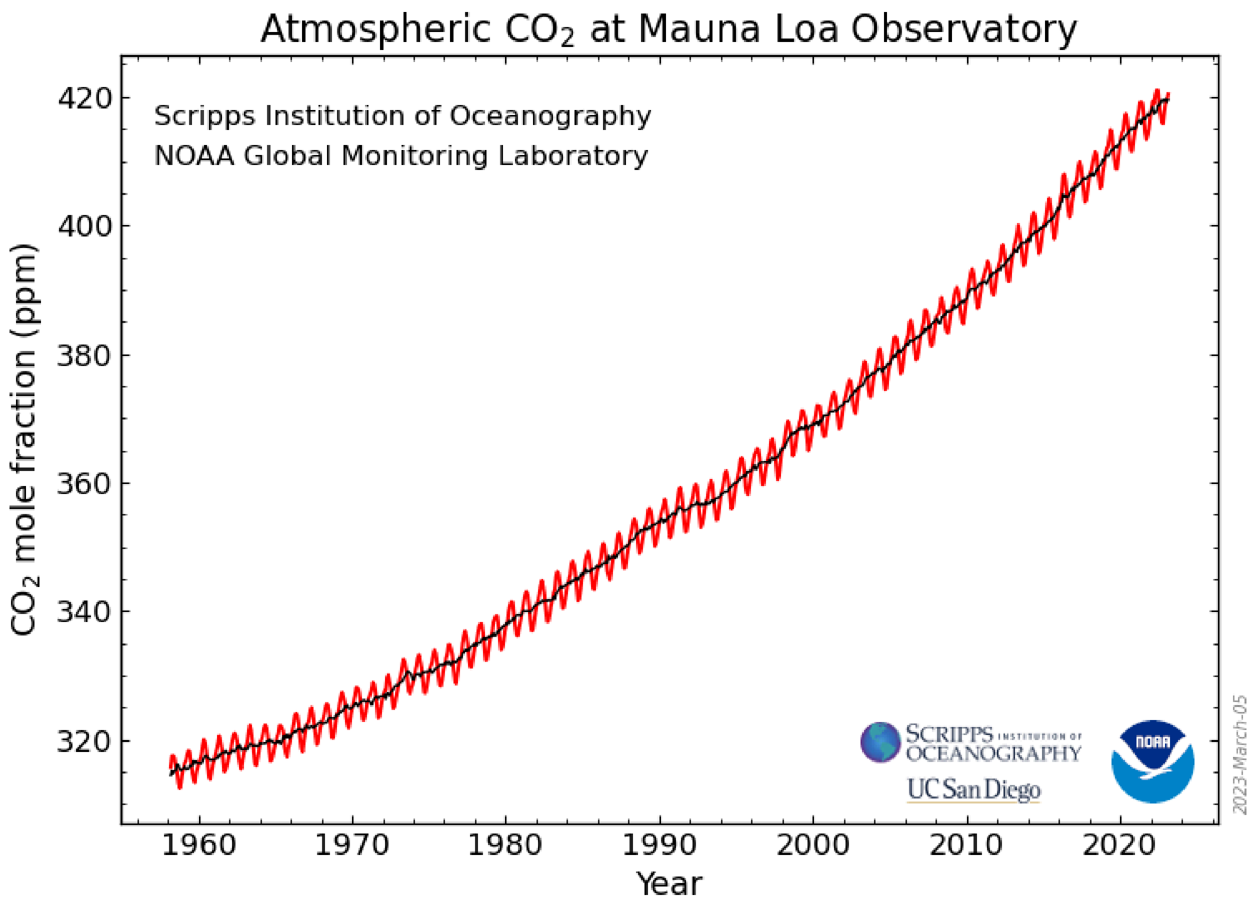


Figure 10: Atmospheric CO_2 measured at Hawaii's Mauna Loa Observatory. The measurements were pioneered by Charles Keeling from Scripps Institution of Oceanography in 1958. From noaa.gov.

Explore Carbon Cycle Data

Goto noaa.gov to explore CO_2 data from other locations. Select a site and then "Carbon Cycle Gases" on the left and then "Timeseries".

- Describe the observations.

Questions

- Ask your parents or grandparents if they have noticed signs of climate changes during their lifetimes. My dad told me that as a boy he used to snow sled down the slopes of a hill in Germany where nowadays there is no more snow.
- By how much did global surface air temperatures increase during the last 100 years?
    - How do we know?
    - What was the average rate of change during that time?
    - What was the rate of change for the last 50 years?
    - Where was the warming larger over the oceans or over land, in the tropics or at high latitudes?
- How much did the upper 2 km ocean heat content change over the last 60 years?
    - How do we know?
    - What was the rate of change?
    - What was the rate of change during the last 20 years?
- How much did the area covered by Arctic sea ice change during the last 35 years?
    - How do we know?
- How have the Greenland and Antarctic Ice sheets changed during the last 15 years?
    - How do we know?
- How much has global mean sea level risen since 1870?
    - How do we know?
    - Calculate the rate of change from 1870 to 2000.
    - Calculate the rate of change from 2000 to the present.
    - Has sea level rise accelerated?
    - Check your answer with this paper.

Videos

Recorded Lecture: Observations Atmosphere & Cryosphere
Recorded Lecture: Observations Oceans & Carbon

References

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Links

https://en.wikipedia.org/wiki/Instrumental_temperature_record
https://www.skepticalscience.com/urban-heat-island-effect-basic.htm

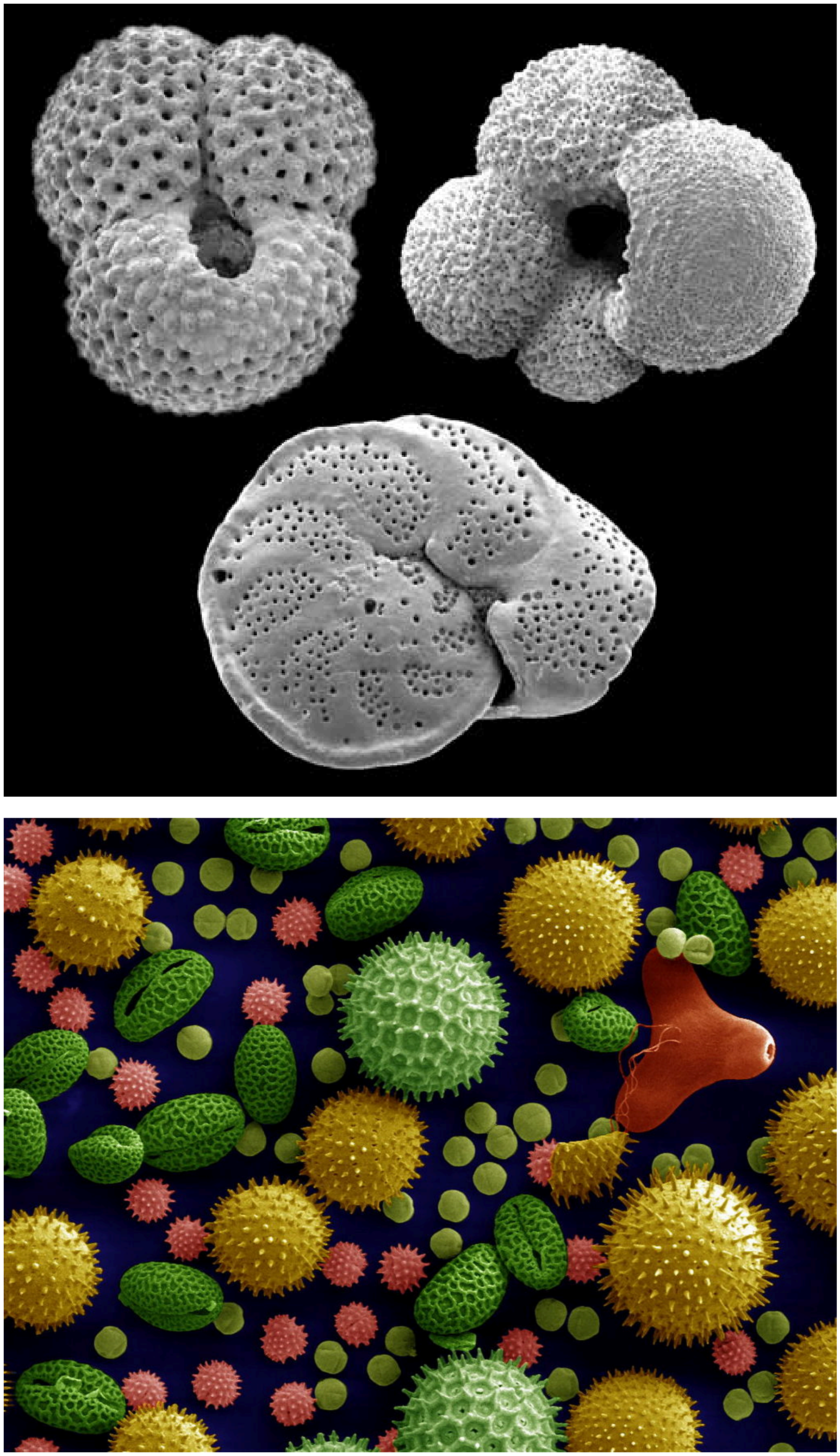
3. Paleoclimate

Measurements with modern instruments (the instrumental record) are available only for roughly the past century. This is insufficient to describe the full natural variability of the climate system, which makes ***attribution*** of observed changes difficult. We want to know if the changes observed in the recent past are unusual compared to pre-industrial climate variability. If they are it is more likely that they are anthropogenic, if not they could well be natural. Paleoclimate research is also important for a fundamental understanding of how the climate system works. Some paleoclimate changes, e.g. the ice age cycles, were much larger than those during the instrumental record. Thus, we can learn much from paleoclimate data about the impacts of large climate changes.

a) Methods

Paleoclimate research is able to extend the instrumental record back in time much further than the instrumental record and has delivered a fascinating history of past climate changes. Most paleoclimate evidence is indirect and based on ***proxies*** for climate variables. This evidence is less precise than measurements with modern instruments because of the additional uncertainty in the relation between the proxy and the climate variable. Examples for proxies are pollen (Fig. 1) found in lake sediments that can be used to reconstruct past vegetation cover, which in turn can be related to temperature and precipitation. Similarly, different species of planktic foraminifera prefer different temperatures. Some live in colder waters others prefer warmer waters. Their fossil shells accumulate in sediments, which can be retrieved with a coring device employed from a research vessel. Shells deeper in the sediment are older. If shells of cold-loving foraminfera are found at a site where currently warm-loving species live, it suggests that near surface temperatures in the past have been colder. Mathematical methods have been developed to quantify the temperature changes from the species composition. Other proxies are chemical such as the ratio of magnesium to calcium (Mg/Ca), which is related to temperature, or isotopes of oxygen or carbon in the calcium carbonate shells of foraminifera, which can be used to reconstruct temperature, salinity, ice volume and carbon cycling. Benthic foraminifera live on or in the ocean's sediments and thus provide useful information on deep ocean properties. Here is an excellent interactive post about paleoclimate proxies.

Figure 1: Electron microscope pictures of fossil foraminifera (top, two planktic foraminifera on top and one benthic foraminifera on the bottom) and pollen (bottom).



Proxies are found in different ***archives*** such as tree-rings, ice-cores, corals, ocean or lake sediment cores that cover different time periods at a range of temporal resolutions (Fig. 2). The ***resolution*** of a record can be quantified as the time difference $\Delta t = t_2 - t_1$ between two adjacent samples t_1 and t_2. The smaller Δt the higher the resolution. Written historical accounts can be used to reconstruct past climatic conditions at very high temporal resolution – some ancient documents contain daily weather entries – back to about 1,000 years, but there are only a limited number of such records available. Tree-rings, corals and speleothems (cave deposits such as stalactites and stalagmites) provide reconstructions at annual to decadal resolution (Δt ~ years to decades) back many thousands of years. Ice cores have typically decadal to centennial resolution going back almost a million years for Antarctica and about 100,000 years for Greenland. Ocean sediment cores cover millions of years in the past but usually at low temporal resolution of centennial to millennial timescales (Δt ~ 100s to 1,000s of years).

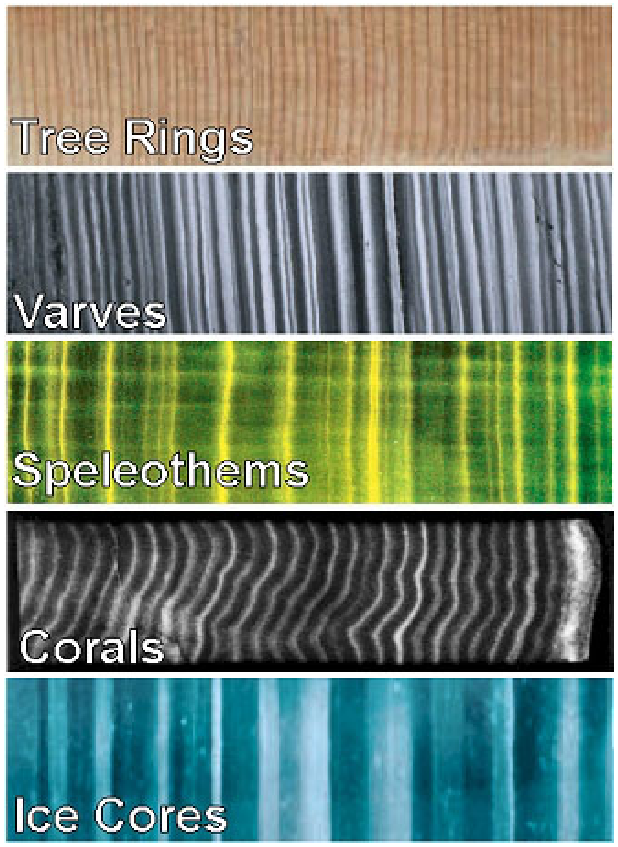


Figure 2: Layered paleoclimate archives. Varves are layered sea or lake sediments.

Several methods are used to date samples to construct ***chronologies*** of paleoclimate records. Tree-rings are annual layers, which can be counted. Patterns of thin and thick rings can be matched from one tree to another, older one (Fig. 3). This way a large number of trees can be used to create a long layer-counted chronology. Layer counting can also be used in other archives with annual layers such as ice cores or lake sediments. Most ocean sediments don't have annual layers because of bioturbation, which is the mixing of sediments by worms and other organisms that live in the sediment. When organic material is present radiocarbon dating can be used to determine the age of a sample. Radiocarbon (^{14}C) decays exponentially with a half-life of 5,730 years. Thus, the lower the ratio of radiocarbon to regular carbon ($^{14}C/^{12}C$) in a sample, the older it is. This ratio can be measured precisely with a mass spectrometer. However, this method can only be used until about 40,000 years before the present because older material has unmeasurably small amounts of ^{14}C.

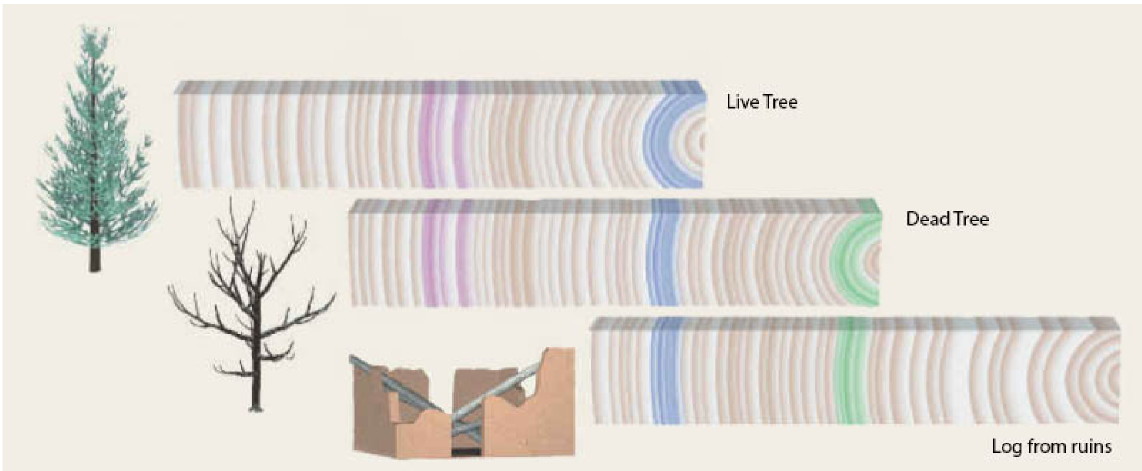


Figure 3: Long tree ring chronologies can be constructed by matching overlapping patterns of different trees.

b) The Last Two Millennia

Historical accounts such as pictures of the frozen Thames (Fig. 4) document a period of relatively cold conditions during the 16th to 19th centuries in Europe called the Little Ice Age. Conversely, relatively warm conditions during the 9th to 13th centuries, called the Medieval Warm Period, may have allowed Vikings to colonize Greenland and travel to North America.

Figure 4: Left: Picture of the frozen Thames from 1683-84 by Thomas Wyke. Right: Ruins of Hvalsey Church from the Greenland settlements of the Norse.

Two recent reconstructions of global temperatures, however, indicate that the Medieval Warm Period was not a global phenomenon (Fig. 5). These reconstructions also suggest that there was a long term cooling trend during the past 2,000 years that culminated in the Little Ice Age, which was terminated by a relatively rapid warming during the 20th century. According to the PAGES 2k reconstruction global average temperature during the three decades from 1971 to 2000 was warmer than at any other 30-year period in the last 1,400 years. This suggests that the recent warming is unusual. The rate of change during the last ~100 years also seems to be unusually fast compared with the previous 2,000 years. The two independent reconstructions agree well in the cooling trend over the past 1,000 years, but the PAGES 2k reconstruction suggests slightly warmer conditions during the first millennium CE (Common Era). The Marcott et al. (2013) dataset is based mostly on lower resolution ocean sediment cores and is therefore smoothed compared to the higher resolution PAGES 2k dataset, which includes mostly land data such as pollen and tree rings.

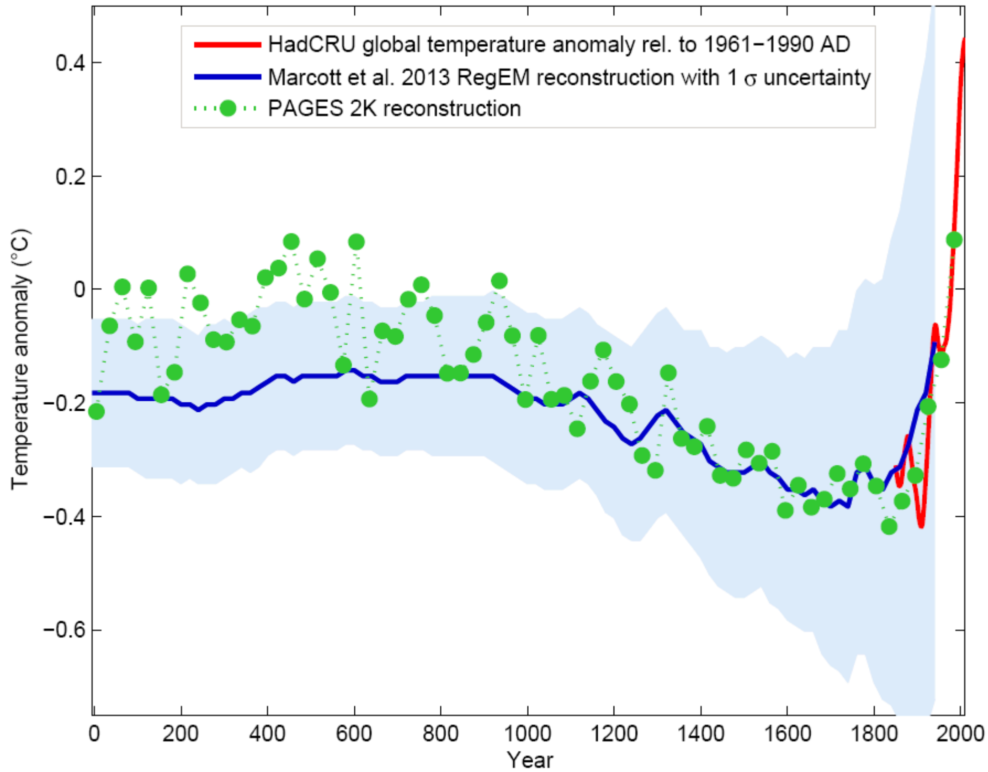


Figure 5: Global multi-proxy temperature reconstructions from the PAGES 2k project based on 7 continental scale regional reconstructions (green) compared with an independent reconstruction that includes the whole Holocene (Marcott et al., 2013; blue with shaded error range; see also Fig. 7) and the instrumental record (red). The PAGES 2k reconstruction represents 30-year averages. It includes many more data than the lower resolution Marcott et al. (2013) reconstruction, particularly for the last 1,000 years. Thus the error range of the PAGES 2k reconstruction is presumably much smaller than that indicated by the blue shading. From RealClimate.

c) The Holocene

Fig. 6 shows the full Holocene (the last 10,000 years) reconstruction of global average temperatures from Marcott et al. (2013). It suggests that the long term cooling trend of the last 2,000 years is part of a longer trend that extends back in time to the middle Holocene around 4,000 BC. The early Holocene from around 8,000 BC to 4,000 BC was relatively warm, similarly to recent decades. (This is debated in the scientific community; a recent paper suggest that it wasn't warmer during the early Holocene and that biases in proxies related to seasonality are to blame. If this is true, the current warming will be unprecedented for more than 10,000 years, perhaps more than 100,000 years or longer.) The rate of temperature change appears to be much smaller compared with the last 100 years, but the relatively low resolution of the reconstruction leads to smoothing and does not allow a fair comparison with the instrumental record on 100 year timescales.

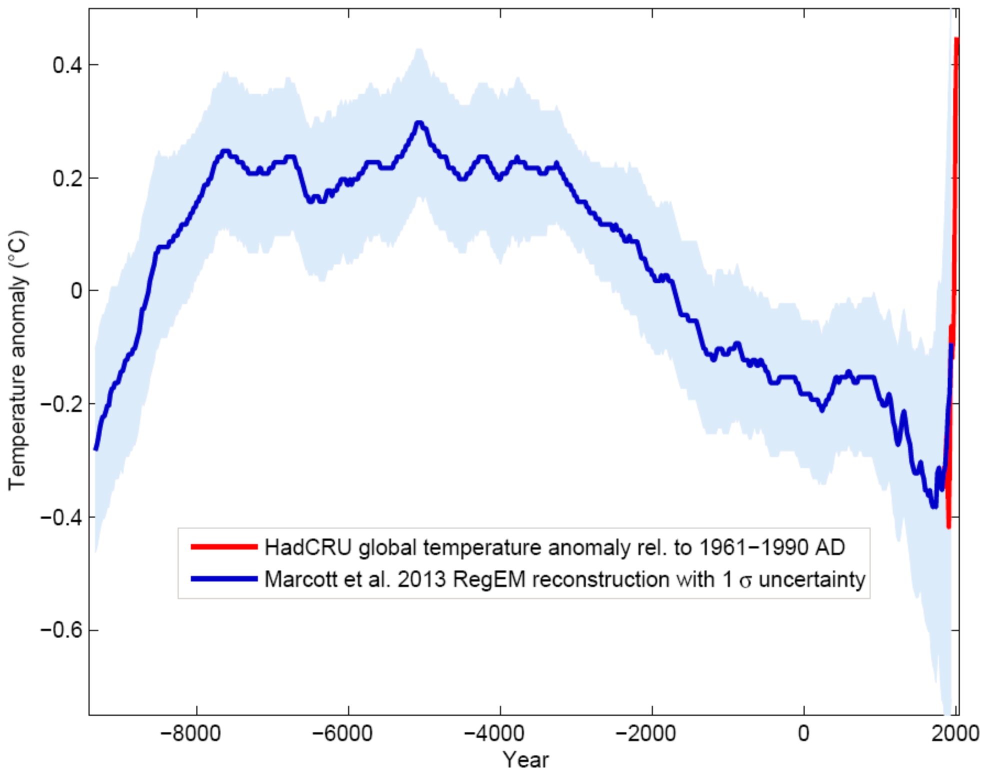


Figure 6: Holocene global surface temperature reconstruction from Marcott et al. (2013, blue) with shaded error ranges together with the instrumental record (red) as a function of time in years CE. From RealClimate.

Now let's have a look at CO_2. Is the observed increase in atmospheric CO_2 during the last 60 years unusual compared to the pre-industrial Holocene? **Ice cores** can be used to answer this question. When snow accumulates on an ice sheet it compresses to firn and later to ice due to the pressure of the overlaying snow (Fig. 7). During this compaction process small bubbles of air are trapped within the ice. In the lab the air can be extracted from the ice, e.g. by mechanically crushing the ice, and its CO_2 concentration, and other greenhouse gases, can be measured.

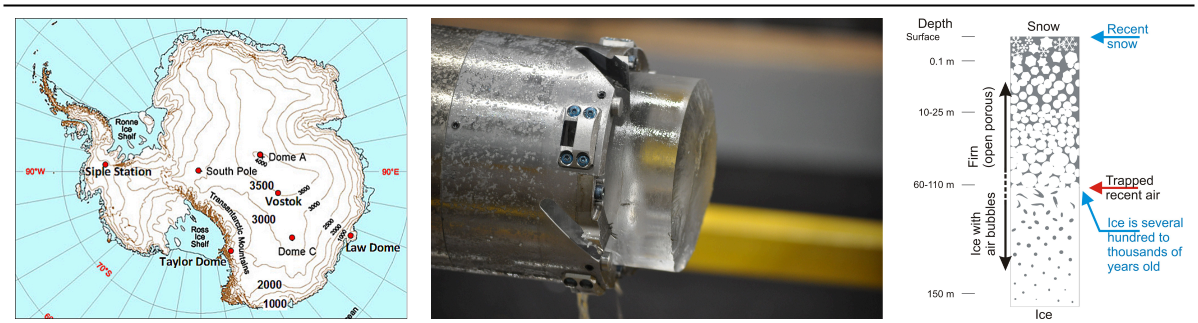
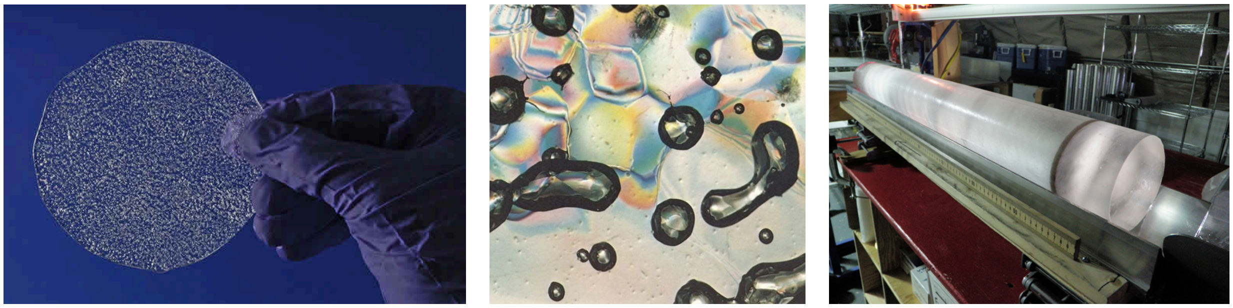


Figure 7: Ice cores have been drilled in different locations in Antarctica (top left) using drilling devices like the one depicted here (top center). Air gets trapped in the ice through compaction of snow and firn (top right). These air bubbles in the ice are visible by eye (bottom left) and in the microscope (bottom center). Sometimes dark ash layers are found in ice cores, which can help to date the ice (bottom right).



Ice cores from Greenland are not suitable for CO_2 reconstructions because they are contaminated by impurities (e.g. dust) that can lead to CO_2 production in the ice. However, Antarctic ice is so pure that it provides excellent records of past atmospheric CO_2 concentrations. Different ice cores have been drilled in Antarctica (Fig. 7). The measurements from the youngest ice and firn match up very well with the direct measurements of modern air from Mauna Loa (Fig. 8). Also, the measurements from different ice cores agree with each other (different colored symbols in Fig. 8). This indicates that Antarctic ice cores faithfully record past atmospheric CO_2 concentrations. The results show that atmospheric CO_2 concentrations have been relatively constant between about 260 and 280 ppm during the Holocene (the last 10,000 years). It was only during the last 200 years that CO_2 concentrations started to increase. Thus, we have answered the question posed above and conclude that the CO_2 increase during the last 200 years is very unusual and has not happened before during the last 10,000 years. We also know that burning of fossil fuels has increased dramatically after the industrial revolution (1760-1840). In the carbon cycle chapter below, more evidence will be presented that demonstrates that the subsequently observed CO_2 increase was indeed due to human activities such as the burning of fossil fuels.

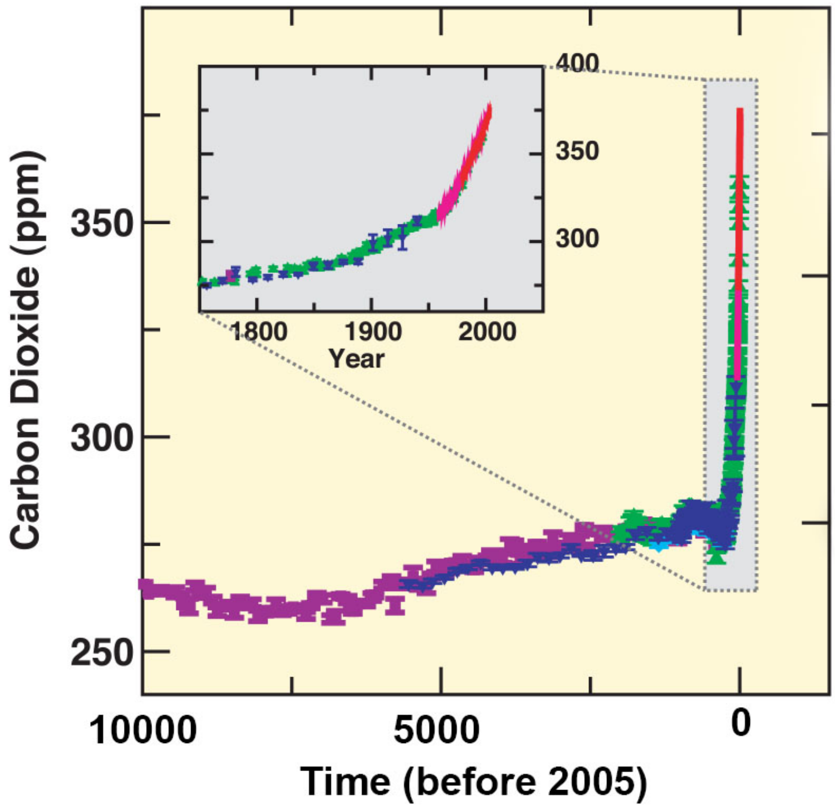


Figure 8: Ice core CO_2 measurements of ancient air. The different colors indicate different ice cores. The inset zooms into the last 200 years and includes in red the modern air measurements from Mauna Loa (see Fig. 8 in Chapter 2). Note that the ice core measurements agree well with the modern data where they overlap. From IPCC (2007).

Other greenhouse gases have also been measured in air extracted from ice cores. Methane (CH_4) and nitrous oxide (N_2O) show a very similar behavior to CO_2, such that their concentrations were relatively constant throughout the Holocene around 700 ppb and 260 ppb, respectively, and increased dramatically during the last 200 years to values around 1,700 and 310 ppb, respectively (IPCC, 2007).

d) The Ice Ages

Fig. 6 already hints at a cold period before the Holocene. Indeed we now know that for a long time Earth had been in an ice age, or glacial state, before the current warm period of the Holocene begun. But it was only in the 19th century that scientists realized that Earth has experienced ice ages on a global scale. This discovery was made by Louis Agassiz, a Swiss geologist, who hypothesized that not only Alpine glaciers were advanced but also large ice sheets moved south from northern Europe and America leaving glacial landforms behind (Fig. 9). For a fascinating and more detailed account of this discovery the reader is referred to Imbrie and Imbrie (1979).

Figure 9: Glacial Landforms. Top: Polished bedrock with striations indicate a glacier was moving over it. The glacier incorporates rocks into its base and by pushing them over the underlying bedrock creates grooves. This example is from Mount Rainier National Park. Bottom left: Moraines (this example is from Svalbard) are glacial deposits formed at the side (lateral moraines) or end (terminal moraines) of a moving glacier. Bottom right: Erratic boulders like this one from Scotland, many miles from a possible bedrock source, have been attributed by Louis Agassiz to the action of ice-age glaciers.

During the height of the last ice age, the Last Glacial Maximum (LGM) roughly 20,000 years ago, large additional ice sheets covered parts of North America and northern Europe (Fig. 10). The Laurentide Ice Sheet was more than 3 km thick and covered all of what is now Canada and part of the northern United States reaching as far south as New York City, Chicago, and Seattle. The Eurasian (or Fennoscandian) Ice Sheet covered all of Scandinavia, much of the British isles, the Baltic Sea, and surrounding land areas from northeastern Germany to northwestern Russia. Mountain glaciers also descended further down valleys and often into the low lands.

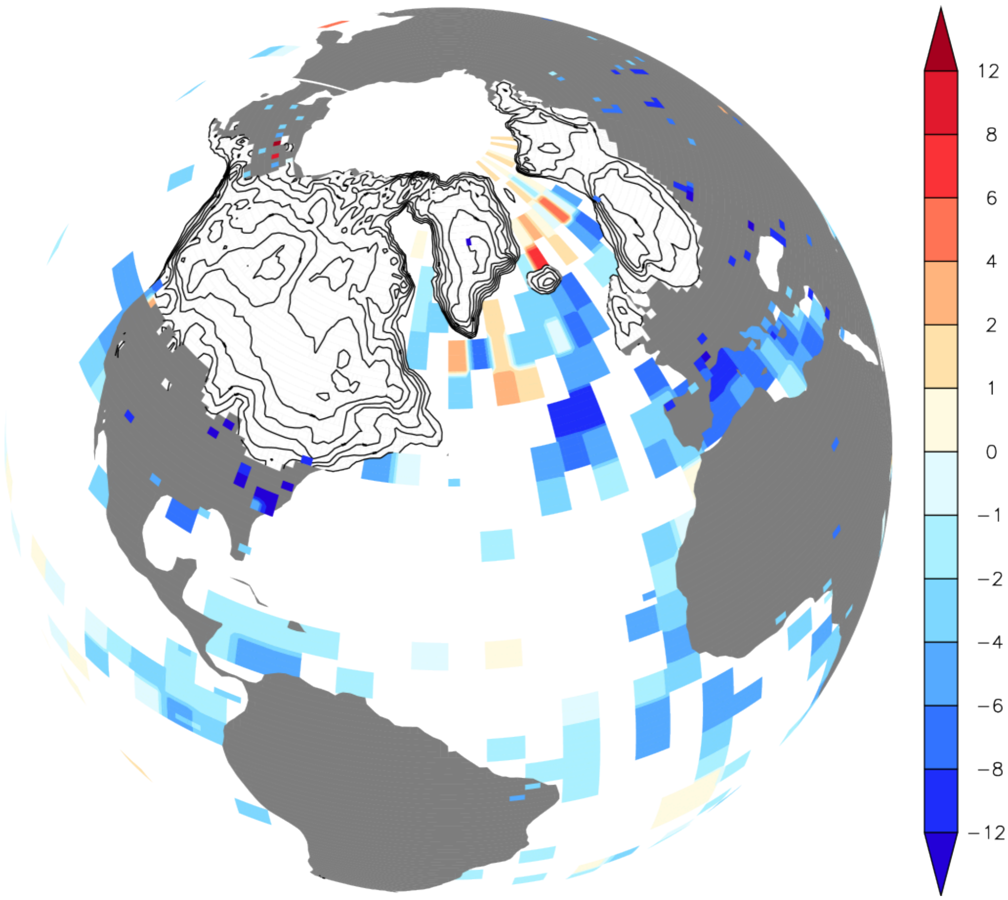


Figure 10: Reconstructions of ice sheets (contour lines show 500 m elevation differences) and surface temperature differences from modern (color scale in K) for the Last Glacial Maximum. From [PAGES news](#).

Because so much more water was locked up as ice on land, sea level was 120 m lower during the LGM than it is today. Imagine your favorite beach of today. There was no water there at the LGM. Explore with [NOAA's interactive bathymetry viewer](#) how much further it would have been to the water during the LGM at your favorite beach.

The LGM is a well-studied time period in paleoclimate research, and we have a wealth of data available. Ice cores show lower concentrations of atmospheric greenhouse gases such as CO_2 (180 ppm vs 280 ppm during the late pre-anthropogenic Holocene; Fig. 11) and methane. Vegetation reconstructions show that forests were replaced by tundra and grasslands over large parts of the mid- to high latitudes (Prentice et al., 2011). We also know from ice and ocean sediment cores that the air was dustier. Temperature proxies show colder temperatures almost everywhere (Fig. 10). However, temperature changes were not the same everywhere. Temperature changes over large parts of the tropical and subtropical oceans were rather small. Globally-averaged sea surface temperatures have been estimated to be only 2°C cooler than the present (MARGO, 2009). Land areas in the tropics experienced moderate cooling of about 3°C (Bartlein et al., 2011). The largest cooling of more than 8°C occurred over land at mid-to-high latitudes and over Antarctica (Fig. 11). Globally averaged surface air temperature has been estimated to be 4°C colder during the LGM (Annan and Hargreaves, 2013). More recent, yet unpublished studies suggest 5°C indicating some uncertainty in these estimates. These authors also suggest that on average the cooling over land was 3 times larger than over the oceans. The ***land-sea contrast*** and ***polar amplification*** are similar to what we've seen in observed warming over the past century (Fig. 2 in Chapter 2). This suggests that those are robust properties of the climate system.

Box 1: Oxygen Isotopes

Isotopes are variations of the same element with a different number of neutrons, which leads to a different mass (Fig. B1). Since different isotopes of the same element have the same number of electrons (yellow circles in Fig. B1) they react chemically identically or very similar.

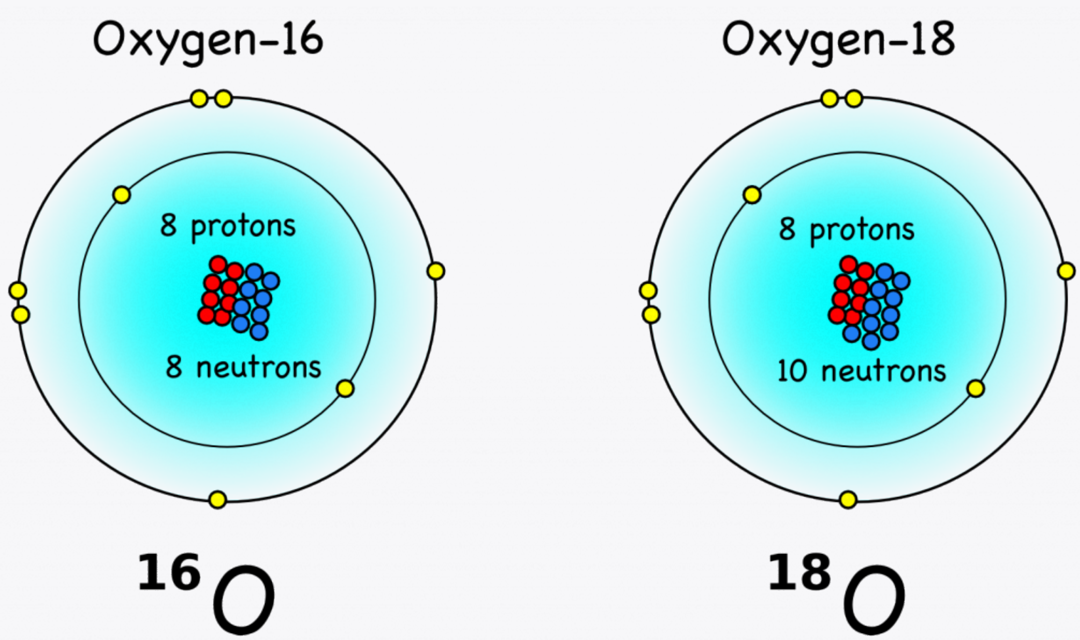


Figure B1: Oxygen's most common (99.8%) isotope oxygen-16 (^{16}O) has 8 protons (red) and 8 neutrons (blue) such that its mass is 16 atomic units. Oxygen-18 (^{18}O) has two additional neutrons, which makes it (18 – 16)/16 = 12.5% heavier than ^{16}O. It is also much rarer (0.2%) than ^{16}O. From [Montessori Muddle](#).

Water molecules (H_2O) with ^{18}O are (20 − 18)/18 = 11% heavier than those with ^{16}O. The mass of a molecule affects how likely it will participate in a phase change such as evaporation or condensation. Water molecules at a certain temperature in the water phase have a distribution of kinetic energy $\frac{1}{2}mv^2$. Some are a little faster, others are a little slower. Only the fastest will be able to leave the water phase and make it to the vapor phase (air). ([Here]() is a nice youtube video explaining this in a little bit more detail.) Because the mass of a heavy water molecule is larger, its velocity, on average, must be smaller in order to have the same kinetic energy. Therefore, the heavier isotopes will remain in the liquid phase more often than the lighter isotopes. This process is called fractionation. It leads to an accumulation of heavy water isotopes in the ocean and relatively more light water isotopes in the air.

Here is an analogy. Imagine a number of black and white soccer balls are lined up at the center line of a soccer field. The black balls are slightly heavier than the white balls. Now a player will shoot the balls, one after another, alternatively black and white, towards the goal. When he/she is done you count the balls that made it across the goal line. Will there be more black or white balls? Yes, indeed, more white balls, because they are lighter and they fly farther due to larger velocities put in by the player who exerts about the same amount of energy for each shot. In this analogy the white balls are the light isotopes.

Isotopes are usually expressed as delta values such as $\delta^{18}O = (R - R_{std})/R_{std}$, where R = $^{18}O/^{16}O$ is the heavy over light ratio of a sample, relative to that of a standard R_{std}. Fig. B2 illustrates how fractionation during *evaporation* and *condensation* affects the isotope values of water, vapor, and ice in the global *hydrological cycle*.

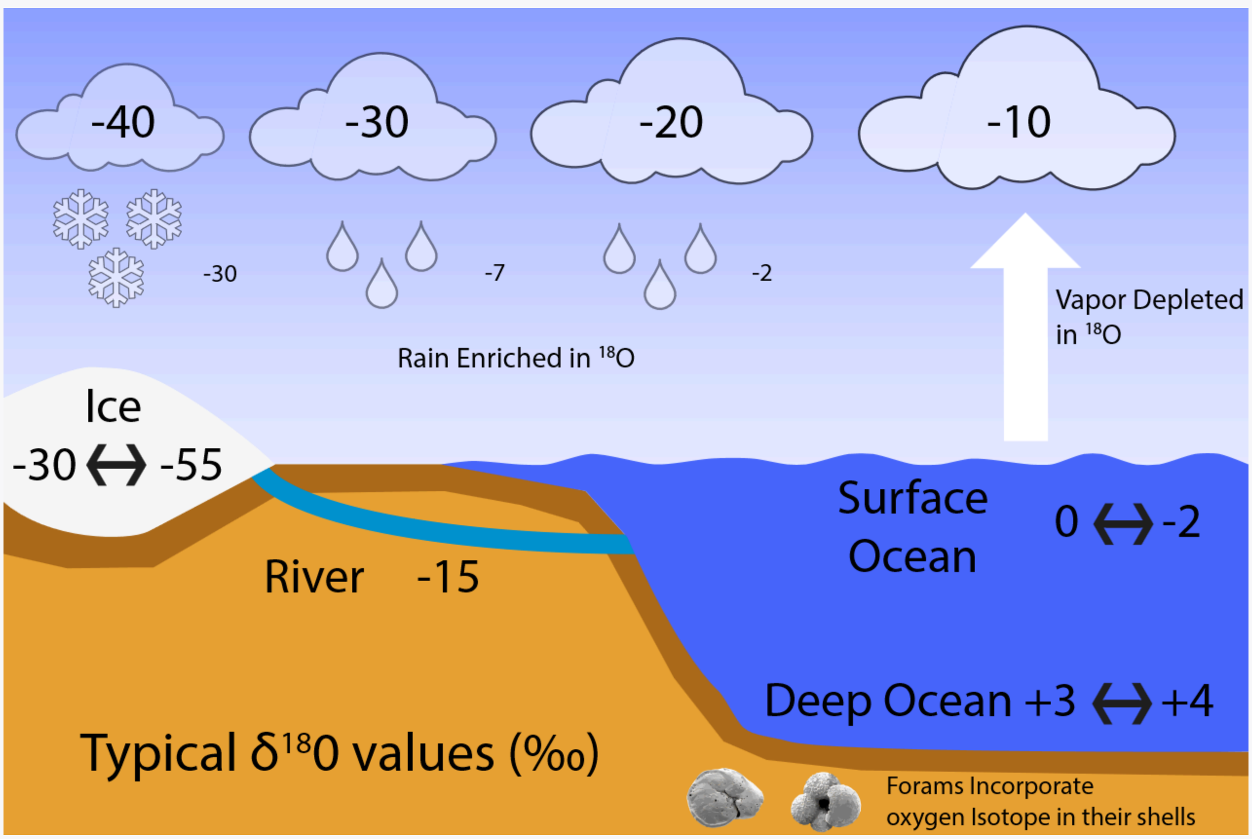


Figure B2: Typical $\delta^{18}O$ values (in permil). Surface ocean water has $\delta^{18}O$ values of around zero. Due to fractionation during evaporation less heavy isotopes make it into the air, which leads to negative delta values of around -10 ‰. Condensation prefers the heavy isotopes for reasons analogous to evaporation. In this example the first precipitation thus has a $\delta^{18}O$ value of about -2 ‰, the remaining water vapor will be further depleted in ^{18}O relative to ^{16}O thus that its $\delta^{18}O$ value is approximately -20 ‰. Any subsequent precipitation event further depletes ^{18}O. This process is known as Rayleigh distillation and leads to very low $\delta^{18}O$ values of less than -30 ‰ for snow falling onto ice sheets. Thus, ice has very negative $\delta^{18}O$ of between -30 and -55 ‰. Deep ocean values today are about +3 to +4 ‰. During the LGM, as more water was locked up in ice sheets, the remaining ocean water became heavier in $\delta^{18}O$ by about 2 ‰. We know this because foraminifera build their calcium carbonate ($CaCO_3$) shells using the surrounding sea water. Thus they incorporate the oxygen isotopic composition of the water into their shells which are preserved in the sediments and can be measured in the lab.

We conclude that paleoclimate data from the LGM show that Earth was dramatically different from today, with large ice sheets, low sea level, and different vegetation. These changes happened even though the global average temperatures changed only by 4-5°C. This is comparable to changes projected for some future scenarios.

Geologic evidence such as that shown in Fig. 9 is abundant only for the last major glaciation because each glacial advance erases evidence from previous glaciations. However, because of our friends, the foraminifera, we know many details of previous glaciations. How can that be? Foraminifera live in the ocean. Well, as explained in the box, the $\delta^{18}O$ of sea water records the amount of ice volume and thus sea level. Since foraminifera record the $\delta^{18}O$ of sea water in their shells, we can amazingly reconstruct past ice volume from tiny shells found in mud on the ocean floor.

Fig. 11 shows that there were about 9 glacial-interglacial cycles during the past 800,000 years. Most of the time sea level was lower than today, during some of the glacial maxima by more than 120 m. Ice core records show that glacial periods were always associated with low atmospheric CO_2 concentrations and low temperatures in Antarctica. CO_2 concentrations varied between about 180 ppm during glacial maxima and 280 ppm during interglacials. The correlation between these completely independent datasets, one from Antarctic ice cores, the other from deep sea foraminifera, is astounding. It demonstrates that climate and the carbon cycle are tightly interlinked. High CO_2 concentrations are always associated with warm temperatures, high sea level, and low ice volume. This indicates the importance of atmospheric CO_2 concentrations for climate but it also suggests that climate impacts the carbon cycle and causes changes in CO_2.

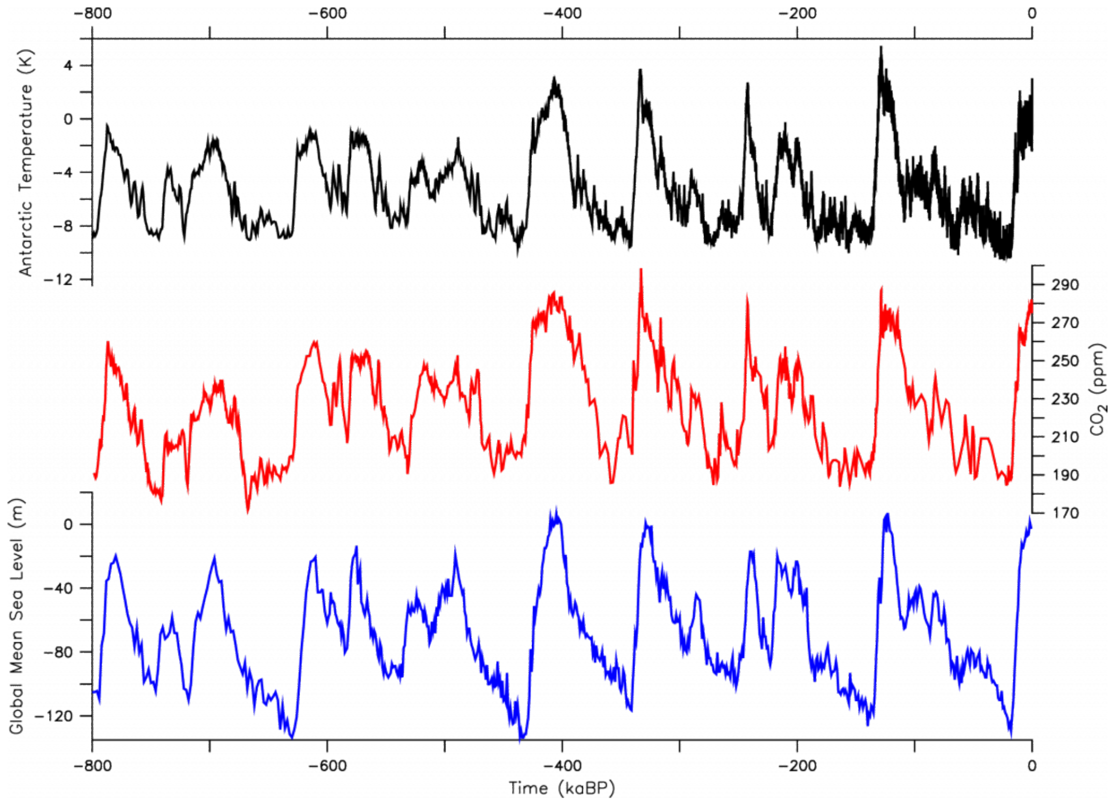


Figure 11: Glacial-interglacial Cycles. Top: Antarctic temperature anomaly (relative to the most recent pre-industrial time) estimated from deuterium ratios measured in the ice (Jouzel et al., 2007). Deuterium (D) is the heavy hydrogen isotope ^{2}H. In H_2O it works similar to the heavy oxygen isotope as a local temperature proxy. Center: Atmospheric CO_2 concentrations from ice cores (Lüthi et al., 2008). Bottom: Global sea level changes (ΔSL = [δ^{18}O − 3.2 ‰]×[-120 m / 1.69 ‰]) estimated from oxygen isotopes (δ^{18}O) measured in benthic foraminifera from deep-sea sediments (Lisiecki and Raymo, 2004). This is a key figure.

Closer inspection of leads and lags shows that during the last deglaciation CO_2 and Antarctic temperature lead global temperature, which suggests that CO_2 is an important forcing mechanism for the warming (Shakun et al., 2012). But we also know that climate affects the carbon cycle. For example, CO_2 is more soluble in colder water, which causes the ice age ocean to take up additional carbon from the atmosphere. Another likely reason for the lower glacial CO_2 concentrations in the atmosphere is iron fertilization. The colder glacial atmosphere was also dustier. Dust contains iron and thus iron delivery to currently iron limited regions such as the Southern Ocean was increased during the ice age. This intensified phytoplankton growth and the biological pump, which describes processes of sinking and sequestering of carbon in the deep ocean. Our recent research indicates that about half (~45 ppm) of the glacial-interglacial CO_2 variations can be explained by temperature and another 25-35 ppm by iron fertilization (Khatiwala et al., 2019). However, this topic is not settled and subject to ongoing research. More about how climate can affect CO_2 will be discussed in the carbon cycle chapter. For now let's just conclude that climate and the carbon cycle are tightly linked.

But what causes glacial-interglacial cycles? They are caused by changes in Earth's orbit around the sun, which affects the seasonal distribution of incoming solar radiation. This theory was proposed in 1938 by Serbian astronomer Milutin Milankovich, who calculated effects of variations in Earth's orbital parameters on solar radiation and linked those to past ice ages. Earth's orbit can be described by three parameters (Fig. 12). The eccentricity E is the deviation from a perfectly circular orbit. Earth's orbit is a slight ellipse although it is close to circular. E varies on ~100,000 year cycles between zero and 0.06. The tilt T, or obliquity, describes the angle between Earth's axis of rotation and the ecliptic, which is the plane of Earth's orbit around the sun. It is currently $T = 23.5°$ but it varies between 24.5° and 22.5° on a 40,000 year cycle. Precession P is the wobble of Earth's axis, like the wobble of a top. Currently we're closest to the sun in January, which corresponds to the axis tilted towards the left in Fig. 12. P varies on a 23,000 year cycle and is strongly modulated by eccentricity. These variations are caused by gravitational forces from the other planets, particularly Jupiter and Saturn.

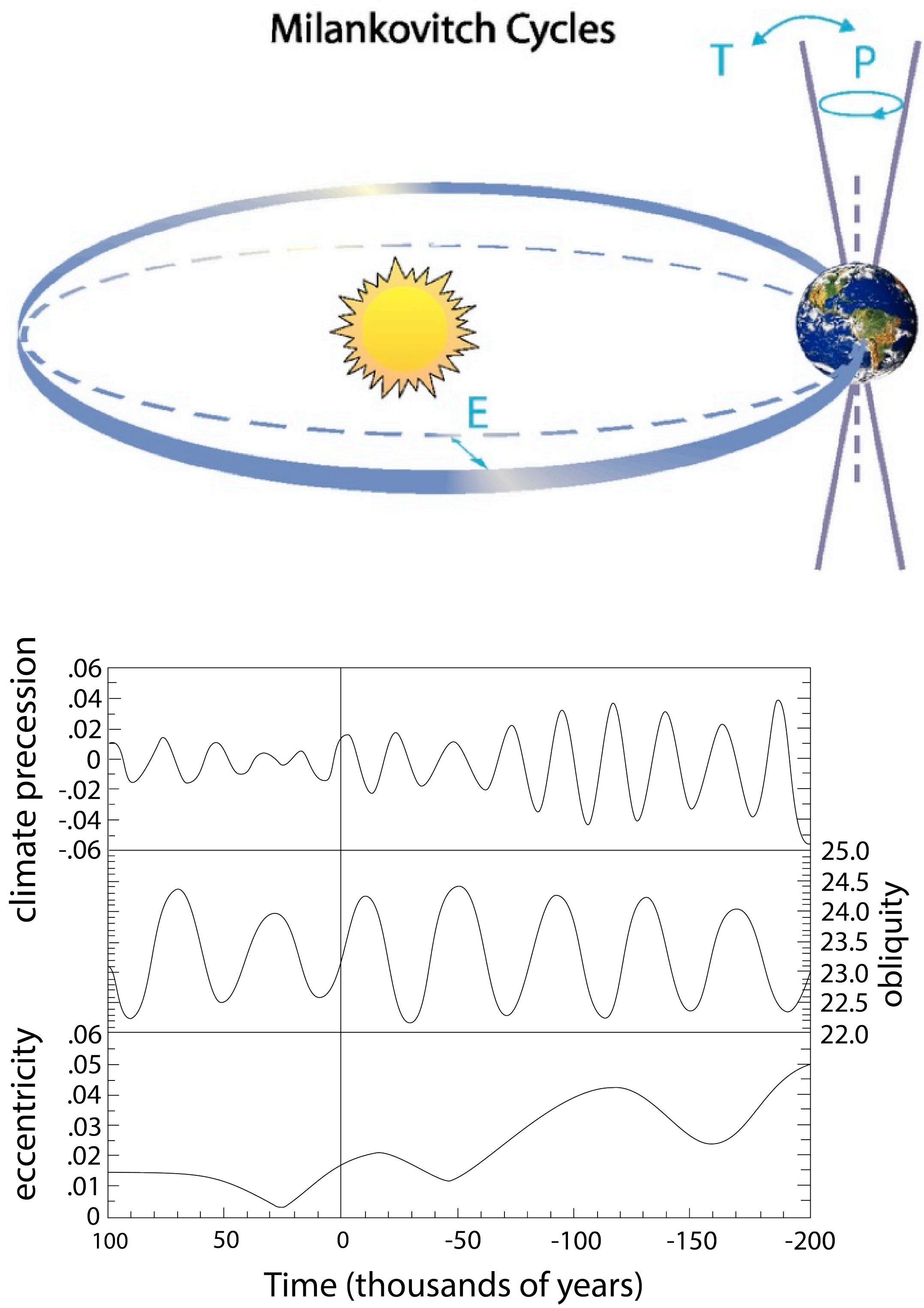


Figure 12: Milankovitch Cycles. Top: Earth's orbit around the sun is determined by eccentricity (E), tilt (T) or obliquity, an precession (P). Bottom: Variation of Earth's orbital parameters through time. Negative numbers towards the left show th and positive numbers show the future.

The idea, proposed by the Irish intellectual Joseph John Murphy and promoted by Milankovitch, Wladimir Köppen and Alfred Wegener, is that summer ***insolation*** in the northern hemisphere controls the waxing and waning of the ice sheets (Berger, 2021). When summer insolation is high, all snow from the previous winter will melt away. When summer insolation is low, some snow survives and during the next winter more snow accumulates. In this way, an ice sheet can grow. The northern hemisphere is important because that is where the major land masses are, where additional ice sheets can grow. The Antarctic ice sheet did not fundamentally change during the ice ages, although it grew somewhat bigger during the ice ages and shrunk a bit during interglacials and the Patagonian ice sheet was much smaller than those in the north.

The orbital forcing theory was essentially confirmed by spectral analysis of deep sea $\delta^{18}O$ data, which shows all the predicted periodicities (Hays et al., 1976). However, exactly when and why ice ages start and end remain active topics of research.

While the orbital forcing theory explains the cyclicity and timing of glacial-interglacial cycles it does not explain their amplitudes (how much global average temperatures have changed). Simulations with global climate models show that the amplitude of glacial-interglacial temperature changes can only be reproduced if CO_2 changes are accounted for (e.g. Shakun et al., 2012). This leads us to conclude that CO_2 changes are an important (feedback) factor in determining glacial-interglacial temperature changes although the ultimate cause of the ice age cycles are Earth's orbital cycles.

Questions

- What are paleoclimate proxies?
- What is a paleoclimate archive?
- Which paleoclimate record has higher resolution: record A, which has data every 100 years, or record B, which as data every 1,000 years?
- What is a chronology?
- Name two methods that are used to date paleoclimate samples?
- How much colder was global average surface air temperatures during the LGM?
- How much colder was global average sea surface temperatures during the LGM?
- How much lower was global sea level during the LGM?
- How much lower was atmospheric CO_2 during the LGM compared to the pre-industrial late Holocene?
- What is the Milankovich theory?
- How do we know the extent of glaciation during the last ice age?
- How do we know sea level and ice volume during previous glaciations?

Videos

Lecture: Last 2,000 Years, Holocene

Lecture: Ice Ages & Isotopes
Lecture: Milankovitch

YouTube: How Ancient Ice Proves Climate Change Is Real from It's Okay To Be Smart

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4. Theory

We have seen how global climate has changed and we've learned that some of these changes have been related to ***forcings*** and ***feedbacks*** such as atmospheric CO_2 concentrations and the seasonal distribution of solar irradiance. Now we want to proceed to understand quantitatively why climate is changing. To do this we will consider Earth's energy budget, review what electromagnetic radiation is, how it interacts with matter, how it passes through the atmosphere, and how this creates the greenhouse effect. But first, we will briefly discuss the general budget equation because it is widely used by scientists and it will be used at different occasions throughout this book.

Box 1: Budget Equation

Scientists like to keep track of things: energy, water, carbon, anything really because it allows them to exploit conservation laws. In physics, for example, we have the law of energy conservation. It is the first law of thermodynamics and states that energy cannot be destroyed or created; it can only change between different forms or flow from one object to another. Similarly, the total amounts of water and carbon on earth are conserved although they may change forms or flow from one component to another one. To start, we need a well-defined quantity of interest. Let's call that quantity X. It could be energy, water, carbon, or something else that obeys a conservation law. It can be restricted to a specific part or component of the climate system, e.g. water in the cryosphere. Mathematically a budget equation can be written as

$$\frac{\partial X}{\partial t} = I - O \ , \tag{B1.1}$$

where the differentials ∂ denote an infinitesimally small change, *t* is time, *I* is the input, and O is the output. The left-hand-side of this equation is the rate-of-change of X. In practice the differentials can be replaced by finite differences (Δ) such that we get

$$\frac{\Delta X}{\Delta t} = I - O \ . \tag{B1.2}$$

Finite differences can be easily calculated: $\Delta X = X_2 - X_1$ and $\Delta t = t_2 - t_1$. Here X_1 corresponds to the quantity X at time t_1 and X_2 correspond to the quantity X at time t_2. Note that the inputs and outputs have units of the quantity X divided by time. They are often called *fluxes*.

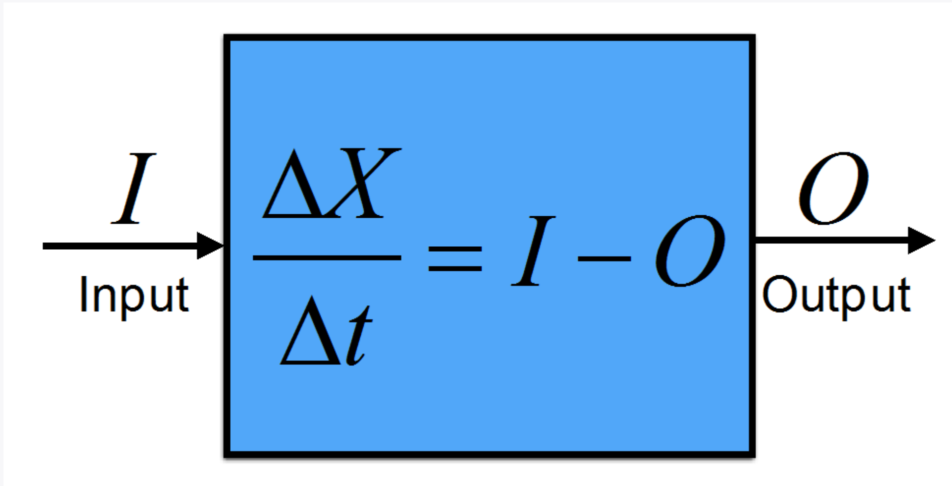


Figure B1.1: Illustration of the budget equation. A box containing a conserved quantity X that does not include internal sources or sinks will change in time according to the external inputs minus the outputs.

Different boxes can be connected such that the output of one box becomes the input of another box.

A budget is in **balance** if the quantity X does not change in time. In this case ΔX = 0 and the input equals the output:

(B1.3) $I = O.$

Let's do a little example. Assume a student gets a monthly stipend of $1,000 and $400 from his/her parents. Those are the inputs to the student's bank account in units of dollars per month I = $1,000/month + $400/month = $1,400/month. The outputs would be the student's monthly expenses. Let's say he/she pays $400 for tuition, $420 for rent, $390 for food, and $100 for books (not for this one though) such that O = $400/month + $420/month + $390/month + $100/month = $1,310/month. The rate-of-change of his/her bank account is ΔX/Δt = I – O = $1,400/month – $1,310/month = $90/month. The student saves $90 per month.

Equation B1.2 can be used to predict the quantity X at time t_2

(B1.4) $X_2 = X_1 + (I - O)\Delta t$

from its value X_1 at time t_1 if the inputs and outputs are known. In climate modeling this method, called forward modeling, is used to predict quantities into the future one time step Δt at a time.

In our example, if the student starts at time t_1, let's say in January, with X_1= $330 in his/her bank account then we can predict that in February he/she will have X_2 = $330 + ($1,400/month – $1,310/month)×(1 month) = $330 + $90 = $420. Note that in this case the time step Δt = 1 month.

a) Electromagnetic Radiation

Earth's energy budget is determined by energy input from the sun (solar radiation) and energy loss to space by thermal or terrestrial *radiation*, which is emitted from Earth itself. Solar radiation has shorter wavelengths than terrestrial radiation because the sun is hotter than Earth. To understand this let's consider the electromagnetic radiation spectrum (Fig. 1). Electromagnetic radiation are waves of electric and magnetic fields that can travel through vacuum and matter (e.g. air) at the speed of light (c). It is one way that energy can be transferred from one place to another. The wavelength of electromagnetic radiation (λ), which is the distance from one peak to the next, varies by more than 16 orders of magnitude. Visible light, which has wavelengths from about 400 nm (nanometers, 1 nm = 10^{-9} m = one millionth of a millimeter) to about 700 nm, occupies only a small part of the entire spectrum of electromagnetic radiation. The frequency (ν) times the wavelength equals the speed of light $c = \nu\lambda$.

Albert Einstein showed in 1905 that electromagnetic radiation has particle properties. In modern quantum physics the light particle is called a photon. Each photon has a discrete amount of energy $E = h\nu = hc/\lambda$ that corresponds to its wavelength, where $h = 6.63\times10^{-34}$ Js is Planck's constant. The shorter the wavelength the higher the energy. High energy photons at ultraviolet, X-ray, and gamma ray wavelengths can be harmful to biological organisms because they can destroy organic molecules such as DNA.

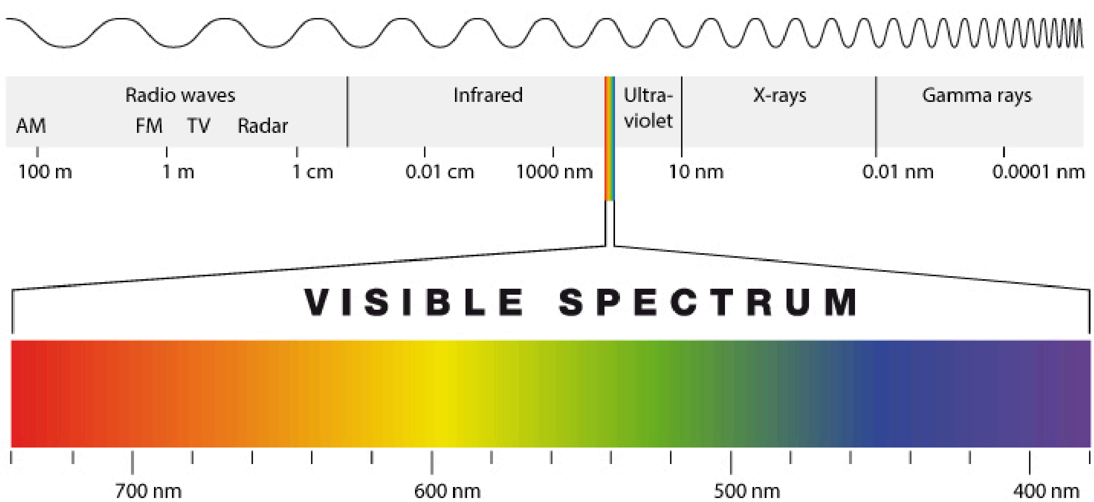


Figure 1: The Spectrum of Electromagnetic Radiation. Electromagnetic radiation ranges from radio-waves with wavelengths of hundreds of meters and more, to gamma rays, with wavelengths of 10^{-12} m, which is as small as the size of an atomic nucleus.

Interaction of electromagnetic radiation with matter depends on the wavelength of the radiation. Molecules have different discrete energy states and they can transition from one state to another one by absorbing or emitting a photon at a wavelength that corresponds to that energy difference (Fig. 2). **Absorption** (capture) of a photon leads to a transition from a lower to a higher energy state. Note that once absorbed the photon is gone and its energy has been added to the molecule. Emission (release) of a photon leads to a transition from a higher to a lower state (reversing the direction of the arrows in the top panel of Fig. 2). Note that the emitted photon can have a different wavelength that the absorbed photon. If, e.g., a UV photon was absorbed and has caused the energy of the molecule to increase from the ground electronic state to the second excited state, the molecule can emit two visible photons, first one that leads to a transition from the second to the first excited electronic state, and then another that leads to a transition to the ground state.

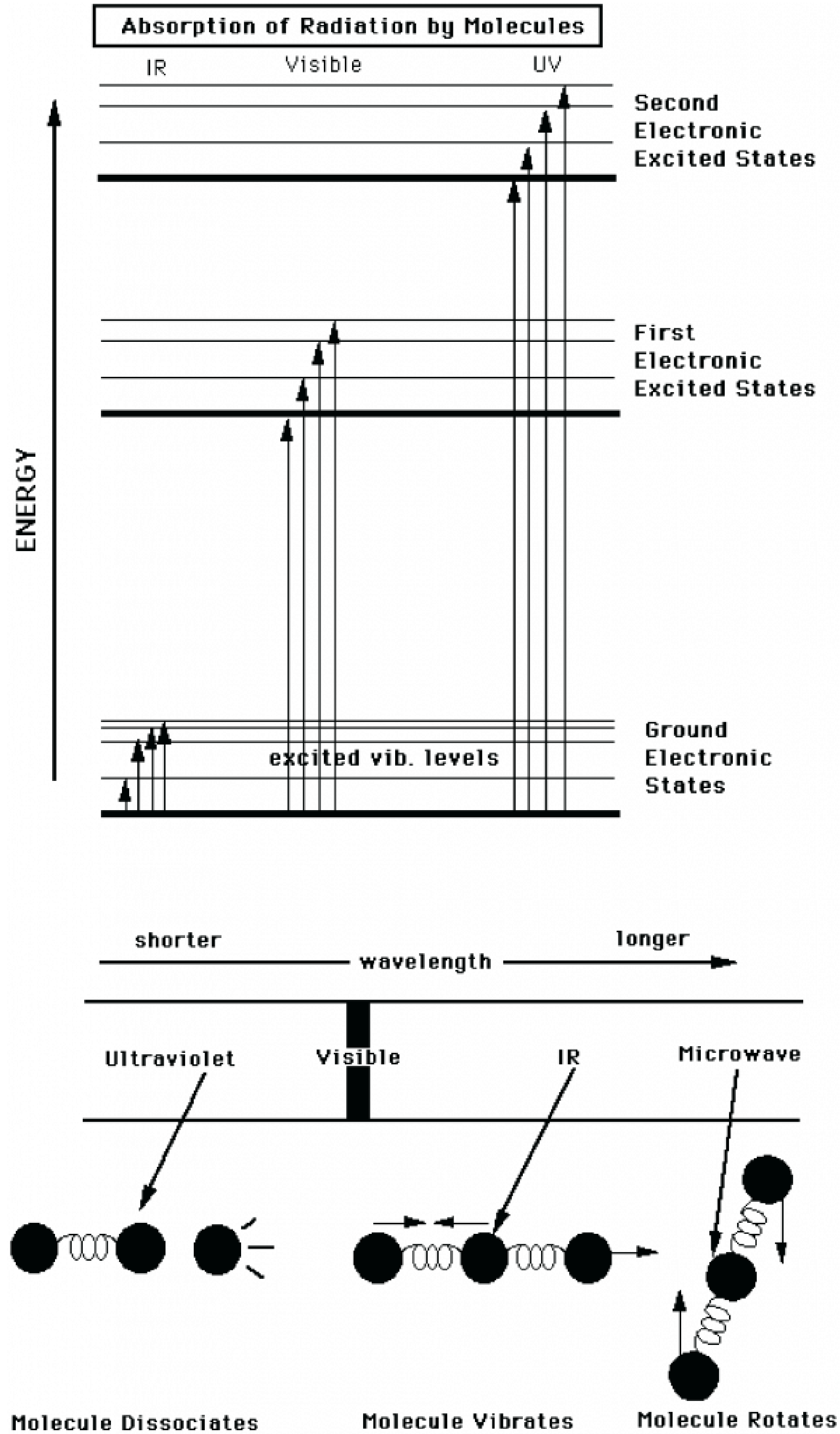


Figure 2: Interactions of electromagnetic radiation with molecules. Top: Absorption of a high-energy photon with an ultraviolet or visible wavelength can lead to excited electronic states. Each excited electronic energy state has sub-states with different vibrational levels. Absorption of a lower energy, infrared photon can excite a vibrational level without changing the electronic state. Bottom: Absorption of low-energy photons at microwave or infrared wavelengths can lead to rotation or vibration of molecules, whereas high-energy photons at ultraviolet wavelengths can break apart molecules. From wag.caltech.edu.

In physics, a ***blackbody*** is an idealized object that can absorb and emit radiation at all frequencies. A blackbody emits radiation according to Planck's law (Fig. 3). In classical physics experiments, a closed box covered on the inside with graphite is used to study its properties. It has only a small hole as an opening to measure the radiation that comes out of the box. Although a blackbody is an idealization many objects behave like a blackbody. Even fresh snow. Or the sun.

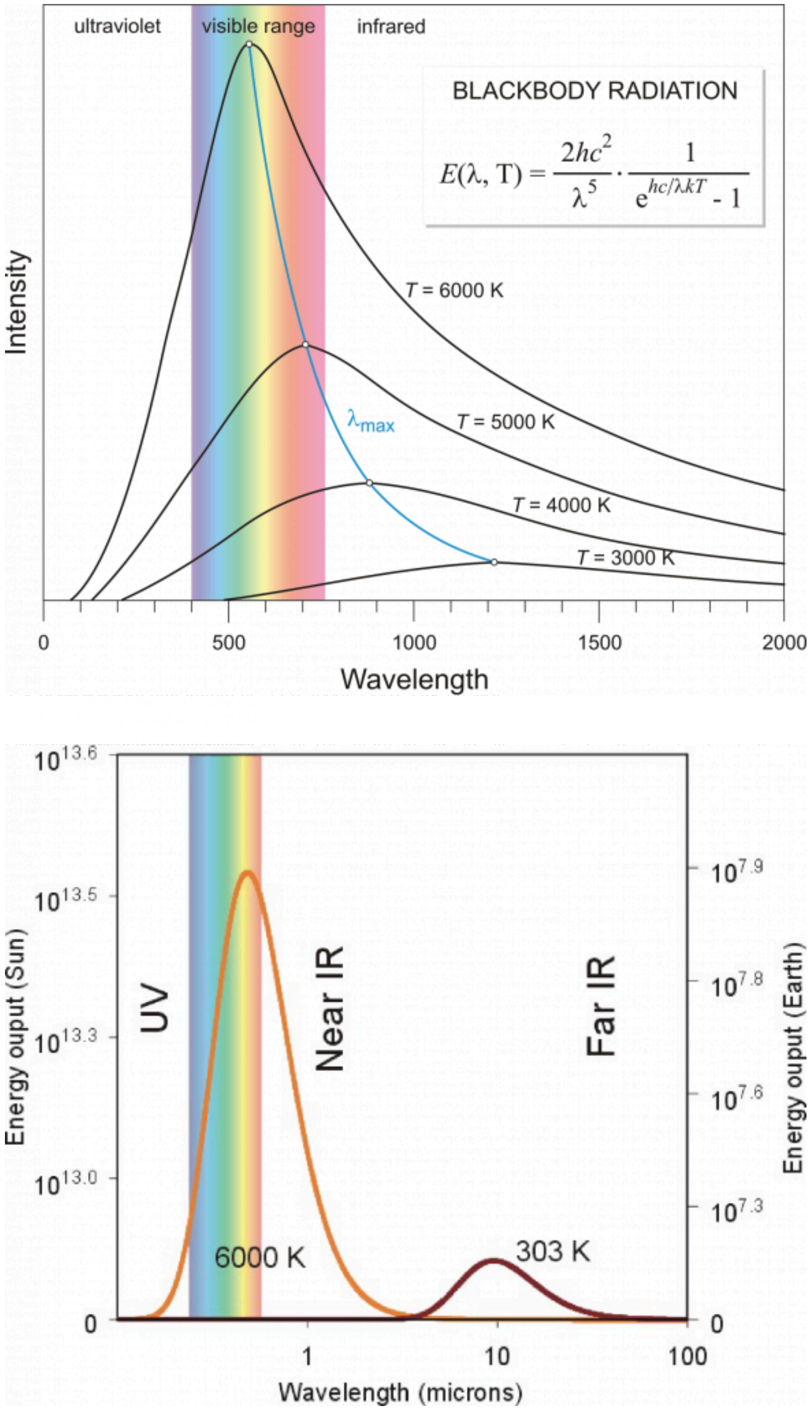


Figure 3: The intensity of blackbody radiation (arbitrary units) according to Planck's law as a function of its wavelength (in nm in the top panel and µm (micrometers, 1 µm = 10^{-6} m = 1,000 nm) in the bottom panel). The sun's temperature is about 6,000 K, with a peak in the visible part of the spectrum. The blue curve in the top panel shows Wien's law, which describes how the maximum moves towards larger wavelengths at colder temperatures. The bottom panel shows curves representative of sun's (6,000 K) and Earth's (303 K) temperatures. Note that the bottom panel uses a logarithmic x-axis and that the values for the sun (left scale) are about 6 orders of magnitude larger than those for Earth (right scale). Top image from [periodni.com](), bottom image from [learningweather.psu.edu]().

Integration of the Planck curve overall frequencies results in the **Stefan-Boltzmann law**

(1) $F = \epsilon\sigma T^4$,

which states that the total energy flux F in units of watts per square meter (Wm^{-2}) emitted from an object is proportional to the absolute temperature of the object *T* in units of Kelvin (K) to the power four. The Stefan Boltzmann constant is σ = 5.67×10^{-8} $Wm^{-2}K^{-4}$ and ε is the emissivity (0 < ε < 1), a material-specific constant that allows for deviations from the ideal blackbody behavior (for which ε = 1). For ε = 1, F represents the area under the Planck curve. The emissivity for ice is 0.97, that for water is 0.96, and that for snow it is between 0.8 and 0.9. Thus, water and ice are almost perfect blackbodies, whereas for snow the approximation is less perfect but still good. Highly reflective materials such as polished silver (ε = 0.02) and aluminum foil (ε = 0.03) have low emissivities.

Equation (1) states that any object at a temperature larger than absolute zero emits energy. The energy emitted increases rapidly with temperature. E.g. a doubling of temperature will cause its radiative energy output to increase by a factor of 2^4 = 16.

Box 2: Earth's Energy Balance Model 1 (Bare Rock)

One or more interactive elements has been excluded from this version of the text. You can view them online here: https://open.oregonstate.education/climatechange/?p=104#video-104-1

Figure B2.1: Illustration of the 'Bare Rock' Energy Balance Model. Yellow arrows indicate solar radiation. The red arrow represents terrestrial radiation.

We can now attempt to construct a simple model of the Earth's energy budget in balance. The energy input is the absorbed solar radiation (ASR). The energy output is the emitted terrestrial radiation (ETR). Thus, equation B1.3 becomes

(B2.1) $ASR = ETR$

The absorbed solar radiation can be calculated from the total solar irradiance (*TSI* = 1,370 Wm^{-2}), which is the flux of solar radiation through a plane perpendicular to the sun's rays. (*TSI* is also sometimes called the solar constant although it is not constant but varies slightly as we'll see below.) Since Earth is a rotating sphere the amount of radiation received per area is S = TSI/4 = 342 Wm^{-2} because the area of a sphere is 4 times the area of a disc with the same radius.

Part of the incident solar radiation is reflected to space by bright surfaces such as clouds or snow. This part is called ***albedo*** (*a*) or reflectivity. Earth's average albedo is about *a* = 0.3. This means that one third of the incident solar radiation is reflected to space and does not contribute to heating the climate system. Therefore ASR = (1 − *a*) S = 240 Wm^{-2}. Assuming Earth is a perfect blackbody ETR = σT^4. With this equation B2.1 becomes

(B2.2) $(1 - a)S = \sigma T^4$

and we can solve for T = ((1 − *a*) S / σ)$^{1/4}$. Inserting the above values for *a*, S, and *σ* gives T = 255 K or T = -18°C, suggesting Earth would completely freeze over as illustrated by ice sheets moving from the pole to the equator in the above animation. This is of course not what is going on in the real world and Earth's actual average surface temperature, which is about 15°C, is much warmer. What's wrong with this model? It is bare rock without an atmosphere! The model works well for planets without or with a very thin atmosphere like Mars, but it fails for planets that have thick atmospheres with gases that absorb infrared radiation such as Venus or Earth.

The concept of Earth's energy balance goes back 200 years to French scientist Jean-Baptiste Fourier, as explained in this (1.5 h) documentary (discussion of Fourier's contributions start at 9:52).

Temperature is the macroscopic expression of the molecular motions in a substance. In any substance such as the ideal gas depicted in Fig. 4 molecules are constantly in motion. They bump into each other and thus exchange energy. A single molecule is sometimes slow and at other times fast, but it is their average velocity that determines the temperature of a gas. More precisely, the temperature of an ideal gas T ~ E is proportional to the average kinetic energy $E = \frac{1}{2}mv^2$ of its molecules. The faster they move the higher the temperature. At absolute zero temperature T = 0 K all motions would cease.

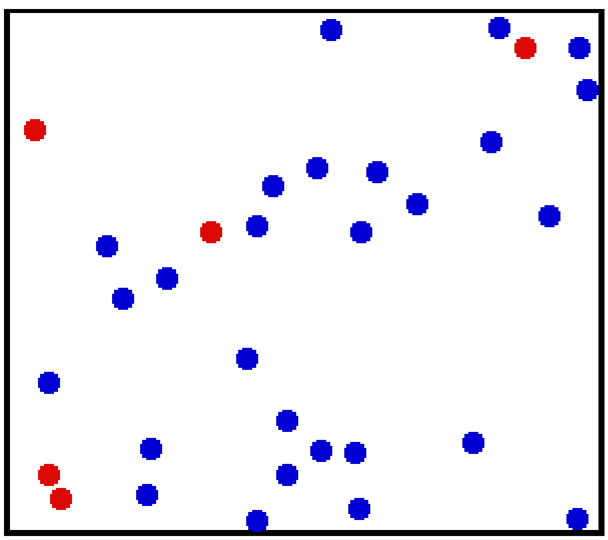


Figure 4: Animation of molecular motions in an ideal gas. From en.wikipedia.org.

The lower panel in Fig. (3) shows blackbody curves for temperatures representative of the sun and Earth. Due to Earth's lower temperature the peak in the radiation occurs at longer wavelengths around 10 µm in the infrared part of the spectrum. The sun's radiation peaks around 0.5 µm in the visible part of the spectrum, but it also emits radiation at ultra-violet and near infrared wavelengths. Sunlight at the top-of-the-atmosphere is almost perfectly described by a blackbody curve (Fig. 5). Some solar radiation is absorbed by gases in the atmosphere but most is transmitted.

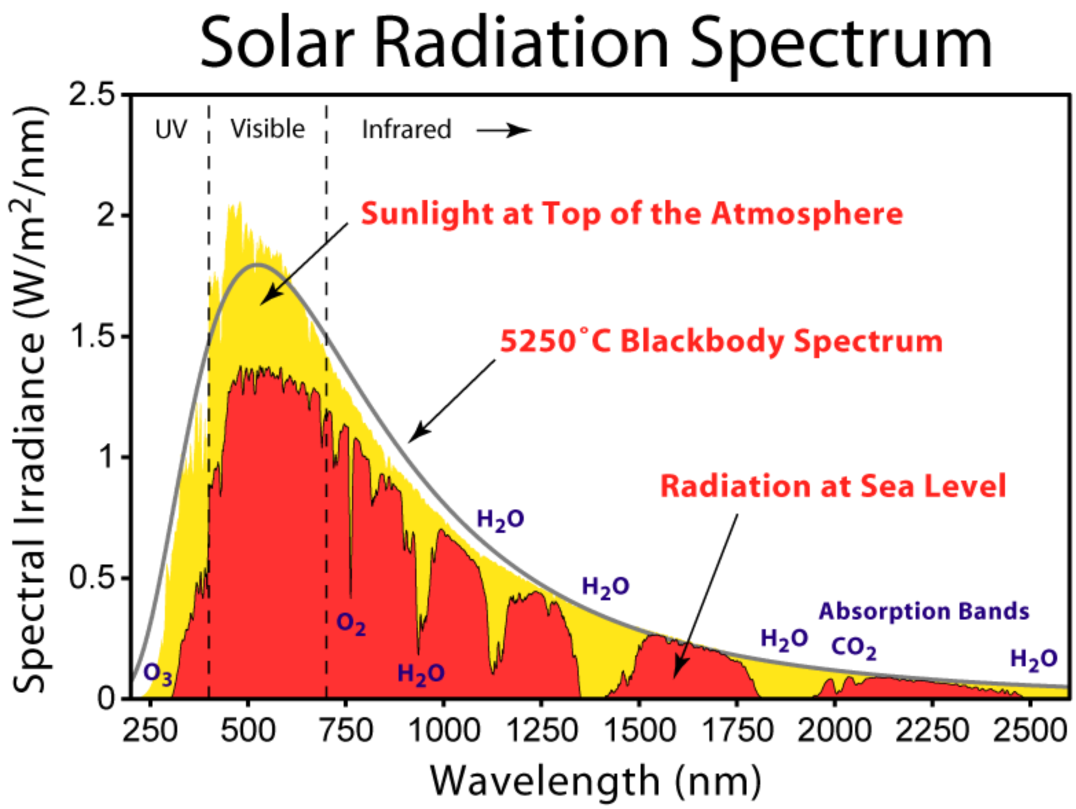


Figure 5. Solar radiation spectra for the incident sunlight at the top-of-the-atmosphere (yellow), at sea level (red), and a blackbody curve (grey). From en.wikipedia.org.

b) The Greenhouse Effect

Absorption by water vapor in the infrared and by ozone (O_3) in the ultraviolet and scattering of light remove 25-30% of solar radiation before it hits the surface (Fig. 6). For Earth's radiation, on the other hand, total absorption is with 70-85% much larger. The most important absorbers in the infrared are water vapor and CO_2 whereas oxygen/ozone, methane, and nitrous oxide absorb smaller amounts. Gases that absorb infrared radiation are called *greenhouse gases*. There is only a relatively narrow window around 10 μm through which Earth's atmosphere allows radiation to pass without much absorption. Thus, Earth's atmosphere is mostly *transparent* to solar radiation, whereas it is mostly *opaque* to terrestrial radiation.

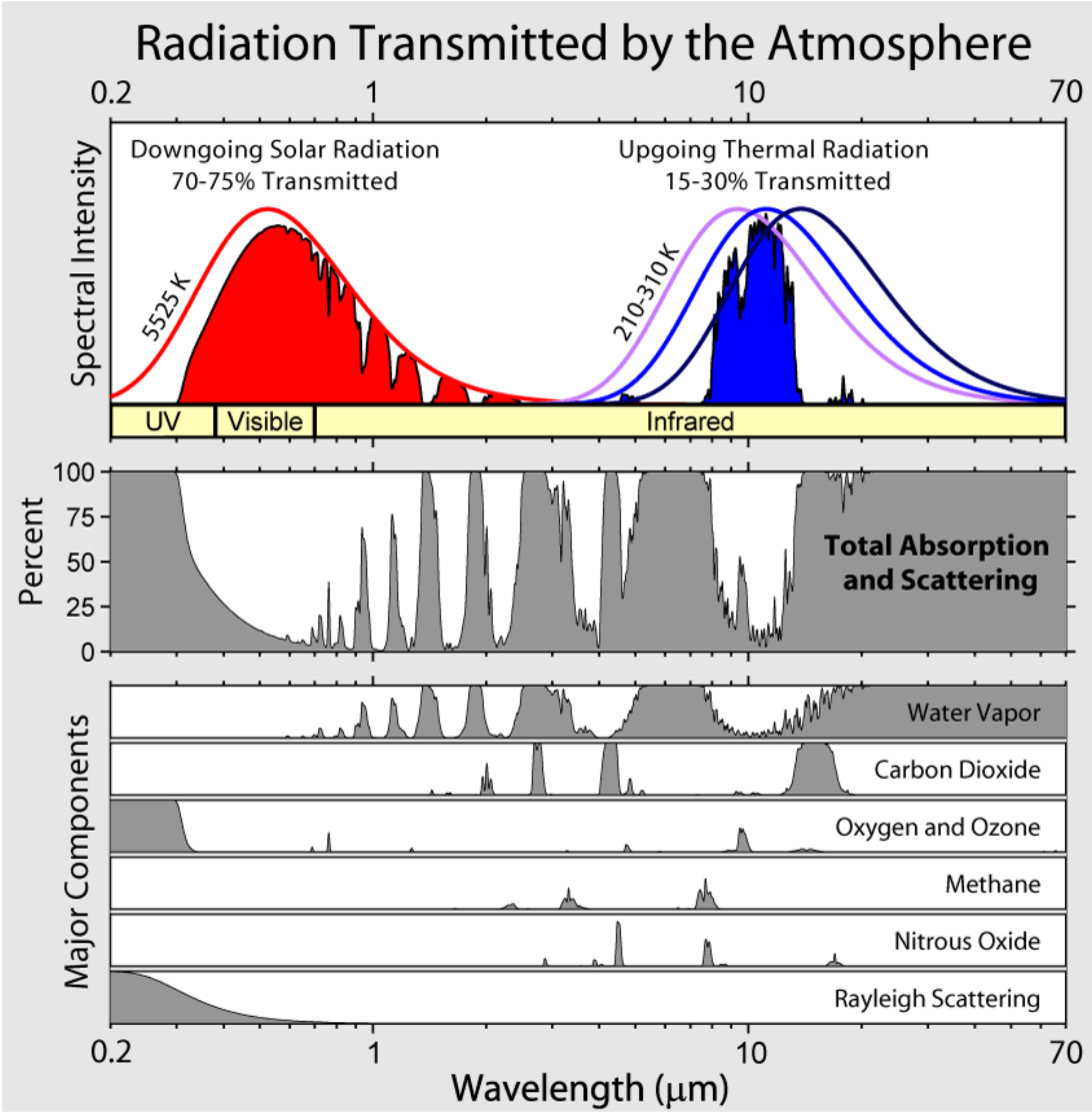


Figure 6: Radiation transmitted and absorbed by the cloud-free atmosphere. The left part of the figure shows the solar radiation and the right part shows Earth's radiation. The blackbody curve at 5525 K (red curve in the top panel) represents the incident (downgoing) solar radiation at the top-of-the-atmosphere. The red filled area is the radiation transmitted through the atmosphere. The difference between the two (the white area between the red curve and the red area) is the amount absorbed by the atmosphere. For Earth's radiation blackbody curves are shown for three temperatures (210, 260, and 310 K) and represent upgoing radiation from the surface. This is a key figure. From commons.wikimedia.org

Why is it that only certain gases in the atmosphere absorb infrared radiation? After all there is much more nitrogen (N_2) and oxygen (O_2) gas in the atmosphere than water vapor and CO_2 (Fig. 7). However, nitrogen and oxygen gas both consist of two atoms of the same element. Therefore, they do not have an electric dipole moment, which is critical for the interaction with electromagnetic radiation. Gas molecules that consist of different elements like water or CO_2, on the other hand, do have dipole moments and can interact with electromagnetic radiation. Since CO_2 is a linear and symmetric molecule it does not have a permanent dipole moment. However, during certain vibrational modes (Fig. 7) it attains a dipole moment and can absorb and emit infrared radiation. Detailed spectroscopic measurements of absorption coefficients show thousands of individual peaks in the spectra for water vapor and CO_2 caused by the interaction of vibrational with rotational modes and broadening of lines by collisions (e.g. Pierrehumbert, 2011). These data are used by detailed, line-by-line radiative transfer models to simulate atmospheric transmission, absorption, and emission of radiation at individual wavelengths.

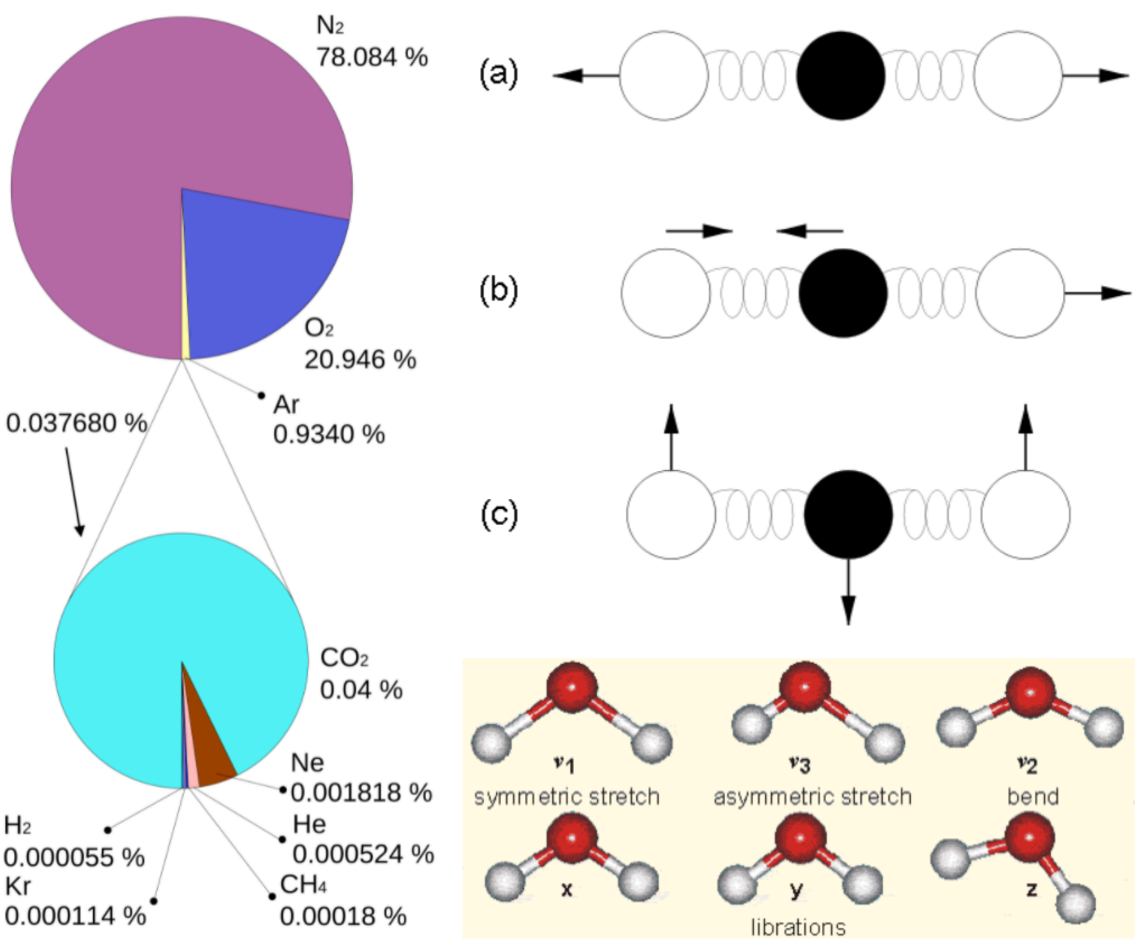


Figure 7:
Left: Composition of the dry atmosphere. Water vapor, which is not included in the image, varies widely but on average makes up about 1 % of the troposphere.
Top right: Vibrational modes of CO_2. The black circles in the center represent the carbon atom, which carries a positive charge, whereas the oxygen atoms (white) carry negative charges. The asymmetric stretch mode (b) and the bend mode (c) lead to an electrical dipole moment, whereas the symmetrical stretch (a) does not. Modes (b) and (c) correspond to absorption peaks around 4 and 15 μm, respectively (Fig. 6).
Bottom right: Vibrational modes of H_2O. The red balls in the center represent the negatively

charged oxygen atom, whereas the white balls represent the positively charged hydrogen atoms. Due to its angle, it has a permanent dipole moment and various modes of vibration and rotation.

Absorption (emission) of radiation by the atmosphere tends to increase (decrease) its temperature. At equilibrium the atmosphere will emit just as much energy as it absorbs, but it will emit radiation in all directions, half of which goes downward and increases the heat flux to the surface. This additional heat flux from the atmosphere warms the surface. This is the greenhouse effect.

An atmosphere in which only radiative heat fluxes are considered and that was perfectly transparent in the visible and perfectly absorbing in the infrared would result in a much warmer surface temperature than our current Earth (see Perfect Greenhouse Model box below). It can also be easily shown that adding more absorbing layers would further increase surface temperatures to $T_s = (n + 1)^{1/4}T_1$, where T_1 = 255 K is the temperature of the top-most of n layers. For two layers T_s = 335 K and the intermediate atmospheric layer's temperature is T_2 = 303 K. This could be called the 'Super Greenhouse Model'. Thus, even though no infrared radiation from the surface can escape to space already with one perfectly absorbing layer, adding more absorbing layers further increases surface temperatures because it insulates the surface further from the top, which will always be at 255 K. In atmospheric sciences, this process is called increasing the optical thickness of the atmosphere.

Box 3: Earth's Energy Balance Model 2 (Perfect Greenhouse)

Since Earth's atmosphere absorbs most terrestrial radiation emitted from the surface, we may want to modify our 'Bare Rock' Energy Balance Model by adding a perfectly absorbing atmosphere. As in the 'Bare Rock' model the energy balance at the top-of-the-atmosphere gives us the emission temperature of the planet, which we now interpret as the atmospheric temperature T_a = 255 K. We now have an additional equation for the atmospheric energy balance. At equilibrium, the total emitted terrestrial radiation from the atmosphere (two times $ETR_a = \sigma T_a^4$ since one ETR_a goes downward and one goes upward) must equal the absorbed radiation coming from the surface ($ETR_s = \sigma T_s^4$). This gives us a surface temperature of $T_s = 2^{1/4}T_a$ = 303 K, which is too warm compared with the real world.

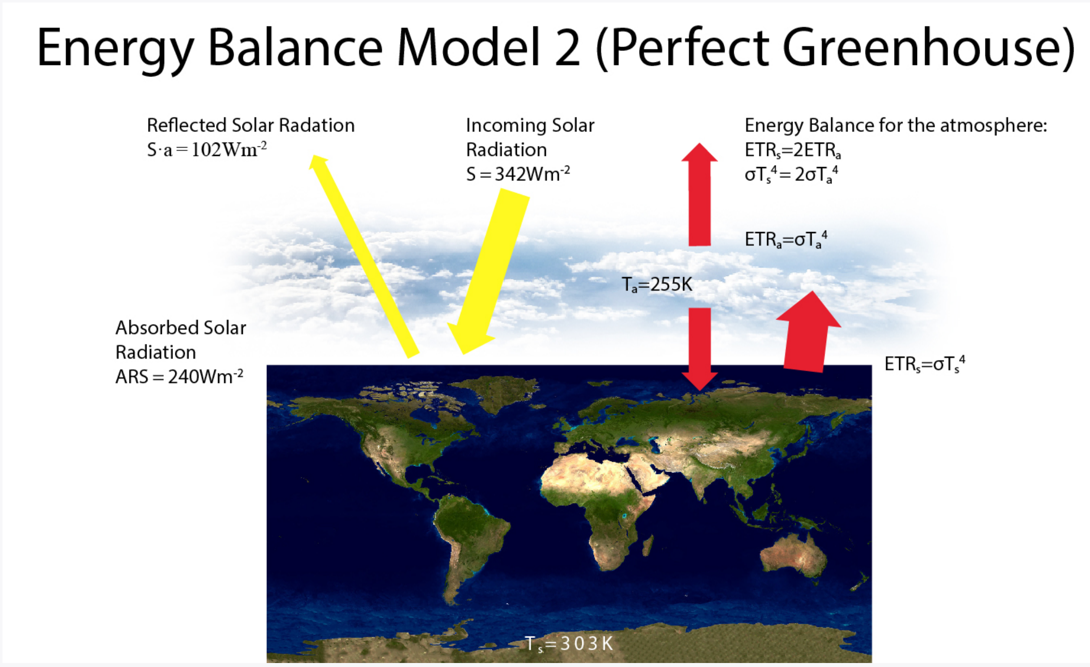


Figure B3.1: As Fig. B2.1 but for the 'Perfect Greenhouse' model.

c) Earth's Energy Budget

In contrast to the 'Perfect Greenhouse' model, Earth's atmosphere does absorb some solar radiation, it does transmit some infrared radiation, and, importantly, it is heated by non-radiative fluxes from the surface (Fig. 8). In fact, if only radiative fluxes (solar and terrestrial) are considered, surface temperatures turn out to be much warmer than they currently are and upper tropospheric temperatures are too cold (Manabe and Strickler, 1964). However, warming of the surface by absorbed solar and terrestrial radiation causes the air near the surface to warm and rise, causing *convection*. Convective motions cause both *sensible* and *latent heat* transfer from the surface to higher levels in the atmosphere. Most of this non-radiative heat transfer is in the form of latent heat. Evaporation cools the surface, whereas condensation warms the atmosphere aloft. Thus, the energy and water cycles on Earth are coupled.

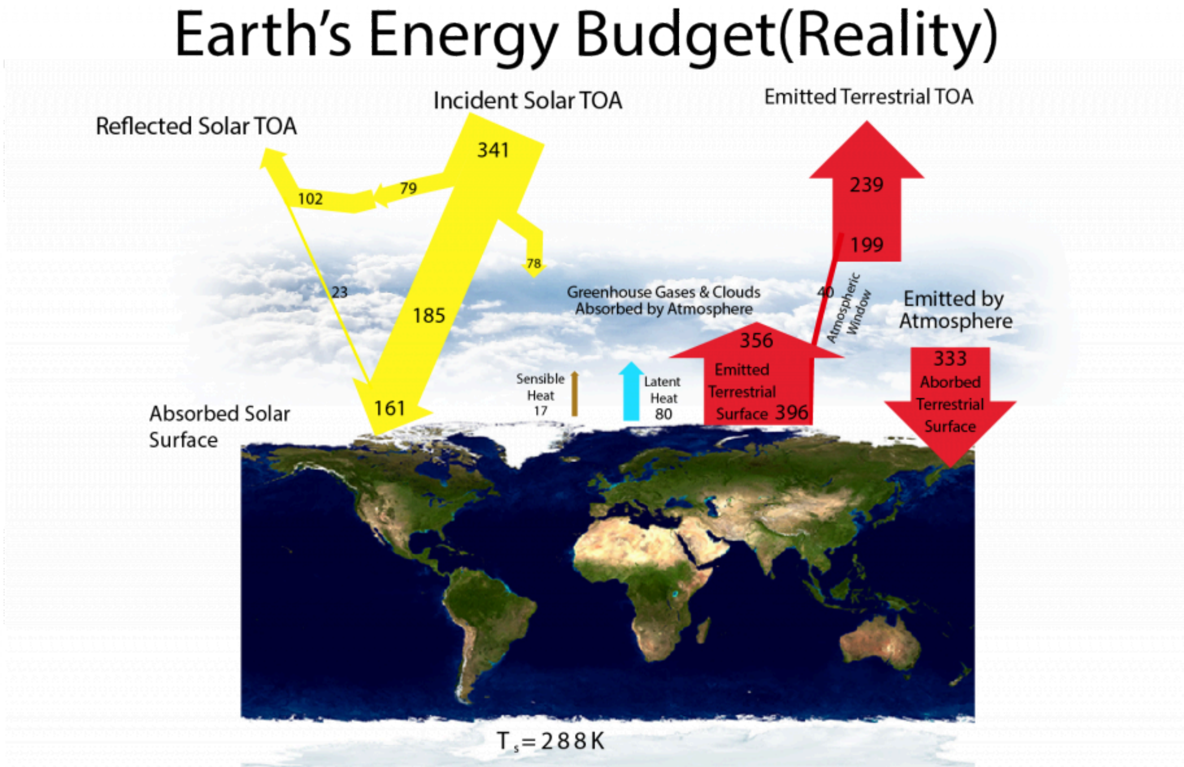


Figure 8: Earth's energy budget estimated from modern observations and models. Adapted from Trenberth et al. (2009).

The downward terrestrial radiation from the atmosphere is the largest input of heat to the surface. In fact, it is more than twice as large as the absorbed solar radiation. This illustrates the important effect of greenhouse gases and clouds on the surface energy budget. The greenhouse effect is like a blanket that keeps us warm at night by reducing the heat loss. Similarly, the glass of a greenhouse keeps temperatures from dropping at night.

Clouds are almost perfect absorbers and emitters of infrared radiation. Therefore, cloudy nights are usually warmer than clear-sky nights. The important effect of water vapor on the greenhouse effect can be experienced by camping in the desert. Night-time temperatures there often get very cold due to the reduced greenhouse effect in the dry, clear desert air.

d) Radiative Forcings, Feedback Processes, and Climate Sensitivity

We've seen how adding greenhouse gases to the atmosphere increases its optical thickness and further insulates the surface from the top, which will lead to warming of the surface. But how much will it warm for a given increase in CO_2 or another greenhouse gas? To answer this question and because we also want to consider other drivers of climate change we introduce the concepts of radiative forcing and feedbacks. These concepts are a way to separate different mechanisms that result in climate change. **Radiative forcing** is the initial response of radiative fluxes at the top-of-the-atmosphere. It can be defined as the change in the radiative balance at the top-of-the-atmosphere (the tropopause) for a given change in one specific process that affects those fluxes with everything else held constant. Examples for such a process are changes in greenhouse gas concentrations, aerosols, or solar irradiance.

A change in the radiative balance at the top-of-the-atmosphere will cause warming if the forcing is positive (more absorbed solar radiation or less emitted terrestrial radiation), and it will cause cooling if the forcing is negative (less absorbed solar or more emitted terrestrial to space). The amount of the resulting warming or cooling will not only depend on the strength of the forcing but also on feedback processes within the climate system. A climate **feedback** is a process that amplifies (positive feedback) or dampens (negative) the initial temperature response to a given forcing. E.g. as a response to increasing CO_2 concentrations surface temperatures will warm, which will cause more evaporation and increased water vapor in the atmosphere. Since water vapor is also a greenhouse gas, this will lead to additional warming. Thus, the water vapor feedback is positive. The warming or cooling resulting from one specific forcing and all feedback processes is called **climate sensitivity**. Let's discuss some of the known radiative forcings and feedback processes in more detail.

Radiative Forcings

Detailed radiative transfer models can be used to calculate the radiative forcing for changes in atmospheric **greenhouse gas concentrations**. As shown in Fig. (9) for **CO_2**, the forcing turns out to depend logarithmically on its concentration (Ramaswamy et al., 2001)

(2) $$\Delta F = 5.35[\mathrm{Wm}^{-2}]\ln(C/C_0),$$

where C is the CO_2 concentration and C_0 is the CO_2 concentration of a reference state (e.g. the pre-industrial).

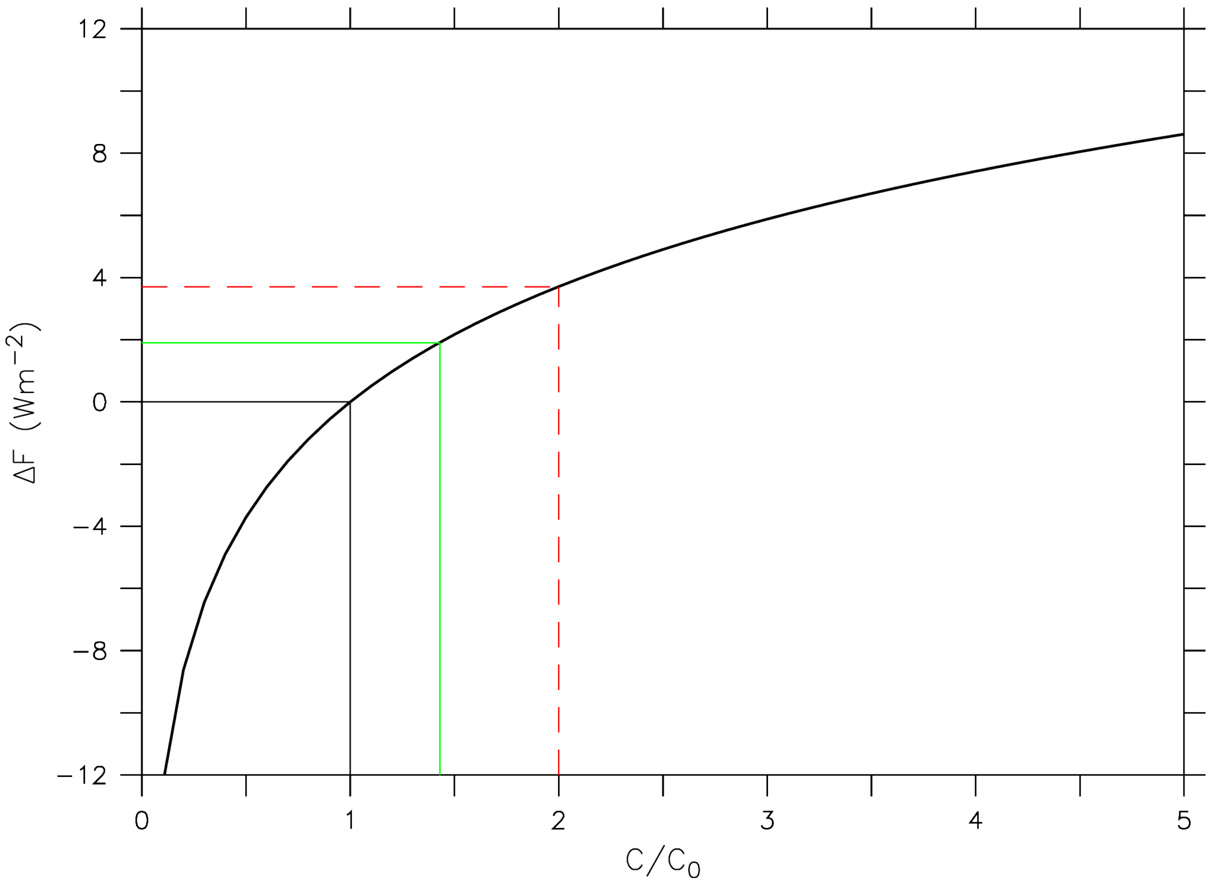
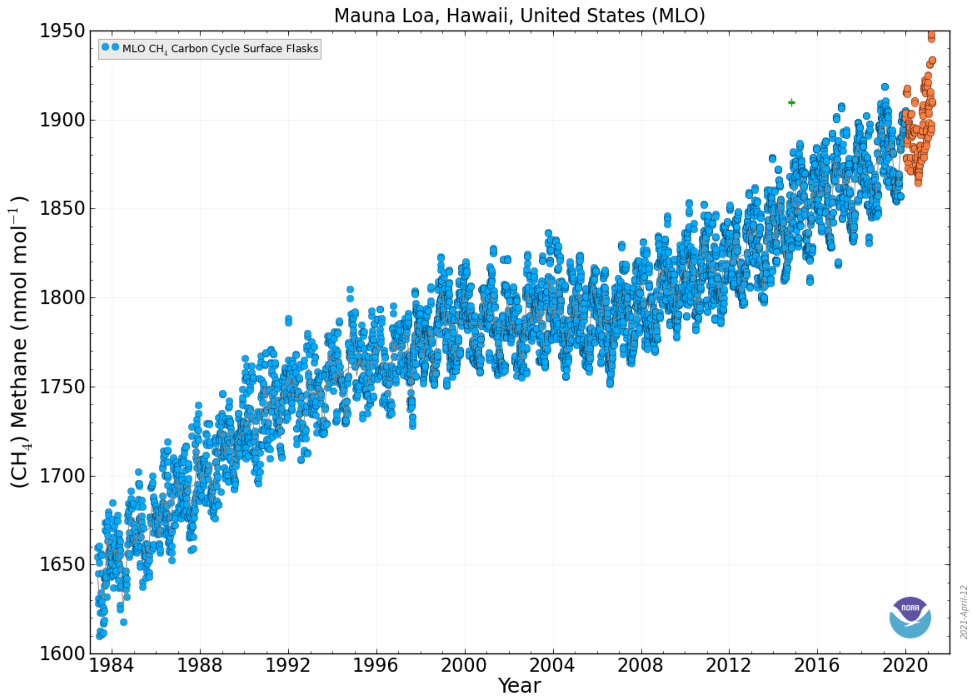
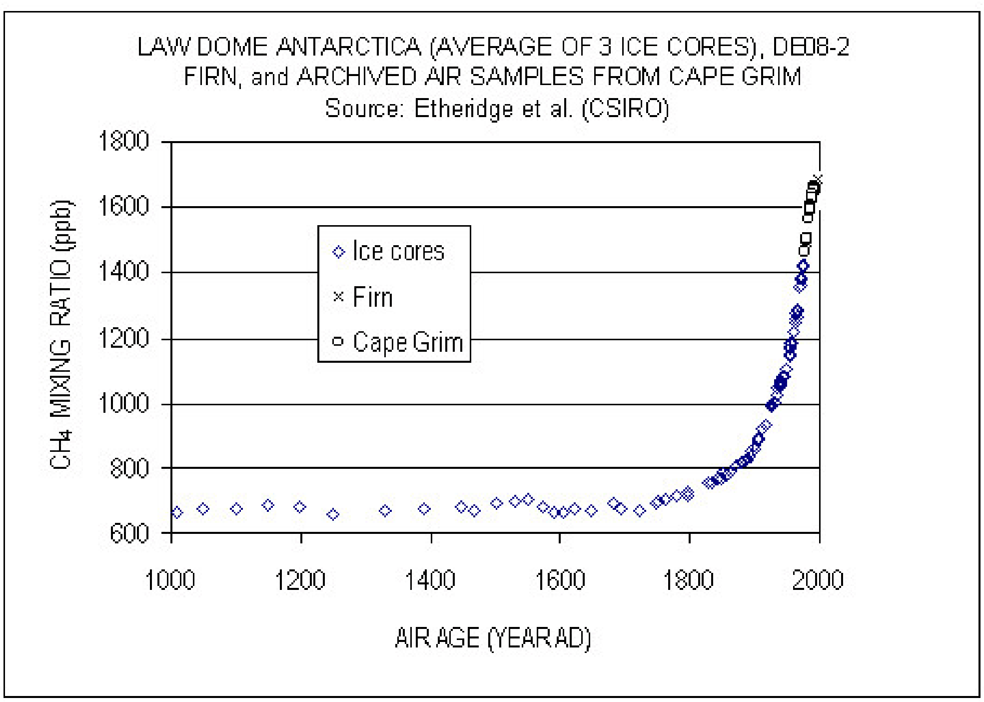


Figure 9: Radiative forcing (ΔF) in watts per square meter as a function of the atmospheric CO_2 concentration (C) relative to a reference value (C_0) according to eq. (2; black thick line). The black straight lines indicate the reference state C = C_0 (ΔF = 0). The green and red dashed lines indicate the current (2016) state relative to the pre-industrial C/C_0 = 400 ppm / 280 ppm = 1.4 (ΔF = 1.9 Wm^{-2}) and that for a doubling of CO_2 (ΔF_{2x} = 3.7 Wm^{-2}) respectively.

This means that the radiative effect of adding a certain amount of CO_2 to the atmosphere will be smaller the more CO_2 is already in the atmosphere. The reason for this is the saturation of peaks in the **absorption** spectrum (Fig. 6). E.g. in the center of the peak at 15 mm all radiation from the surface is already fully absorbed. Increasing CO_2 further only broadens the width of the peak.

Figure 10: [Top:](#) Atmospheric methane concentrations as a function of time. Top: Recent air measurements from Mauna Loa.
[Bottom:](#) Ice core, firn, and air measurements from Antarctica. Note that most methane sources are in the northern hemisphere, which leads to higher concentrations there compared with the southern hemisphere.



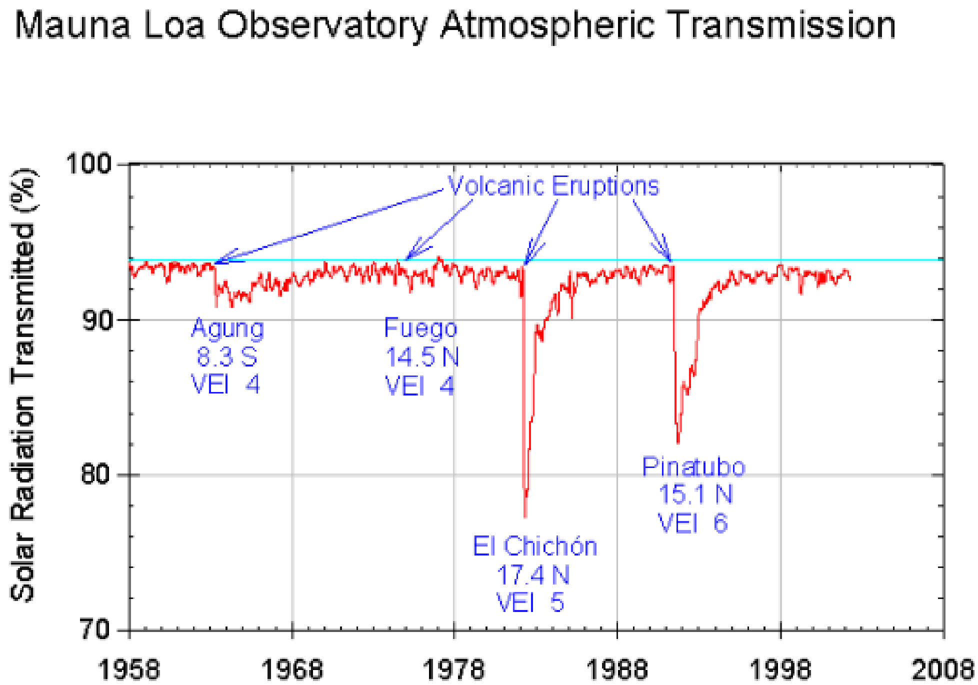


Methane (CH_4) is produced naturally in wetlands and by various human activities such as in the energy industry, in rice paddies, and by agriculture (e.g. cows). It is removed by chemical reaction with OH radicals and has a lifetime of about 10 years. Anthropogenic activities have increased atmospheric methane concentrations since the industrial revolution by more than a factor of two, from around 700 ppb to more than 1600 ppb currently (Fig. 10). On a per-molecule basis methane is a much more potent greenhouse gas than CO_2, perhaps in part because its main absorption peak around 8 mm is less saturated than the one for CO_2 (Fig. 6). However, since methane concentrations are more than two orders of magnitude smaller than CO_2 concentrations, its radiative forcing since the industrial revolution is 0.5 Wm^{-2}, which is less than that for CO_2. As we will see below CO_2 also has a much longer lifetime than methane and therefore it can accumulate over long timescales. Indeed, while recent measurements indicate a slowdown of methane growth rates in the atmosphere, CO_2 increases at ever higher rates (Fig. 8 in Chapter 2).

Aerosols are small particles suspended in the air. Natural processes that deliver aerosols into the atmosphere are dust storms and volcanic eruptions. Burning of oil and coal by humans also releases aerosols into the atmosphere. Aerosols have two main effects on Earth's radiative balance. They directly reflect sunlight back to space (direct effect). They also act as cloud condensation nuclei such that they can cause more or brighter clouds, which also reflect more solar radiation back to space. Thus, both the direct and indirect effects of increased aerosols lead to cooling of the surface. Therefore, aerosol forcing is negative.

Large explosive volcanic eruptions can lead to ash particles and gas such as sulfur dioxide (SO_2) ejected into the stratosphere, where they can be rapidly distributed over large areas (Fig. 11). In the stratosphere SO_2 is oxidized to form sulfuric acid aerosols. However, stratospheric aerosols eventually get mixed back into the troposphere and removed through precipitation or dry deposition. The lifetime of volcanic aerosols in the stratosphere is in the order of months to a few years. Estimates of the radiative forcing from volcanic eruptions depend on the eruption but varies from a few negative tenths of a watt per square meter to -3 or -4 Wm^{-2} for the largest eruptions during the last 100 years (Fig. 12).

Figure 11: Effects of volcanic eruptions.
[Top:](#) Measurements of solar radiation transmitted at Hawaii's Mauna Loa Observatory.
[Center:](#) Photograph of a rising Pinatubo ash plume.
[Bottom:](#) Photograph from Space Shuttle over South America taken on Aug. 8, 1991 showing the dispersion of the aerosols from the Pinatubo eruption in two layers in the stratosphere above the top of cumulonimbus clouds.



Anthropogenic aerosols are released from burning tropical forests and fossil fuels. The latter is the major component producing more sulfate aerosols currently than those naturally produced. Aerosol concentrations are higher in the northern hemisphere, where industrial activity is located. Radiative forcing estimates are uncertain but vary between about -0.5 to -1.5 Wm^{-2} for both direct and indirect effects (Figs. 12, 13). Generally, estimates of aerosol forcings are more uncertain than those for greenhouse gases.

Solar irradiance varies with the 11-year sunspot cycle. Direct, satellite-based observations of total solar irradiance show variations of about 1 Wm^{-2} between sunspot maxima and minima (Fig. 14). To estimate radiative forcing TSI needs to be divided by four, which results in about 0.25 Wm^{-2}. Longer-term estimates of TSI variations based on sunspot cycles indicate an increase from the Maunder Minimum (1645-1715) to the present by about 1 Wm^{-2}. The resulting forcing is again about 0.25 Wm^{-2}.

Comparisons of the different forcings indicate that long-term trends of the last 100 years or so are dominated by anthropogenic forcings. The negative forcings from increases in aerosols compensate somewhat the positive forcings from the increase in greenhouse gases. However, the net effect is still a positive forcing of about 2 Wm^{-2}. Volcanic forcings are episodic and the estimates of solar forcing are much smaller than those of anthropogenic forcings.

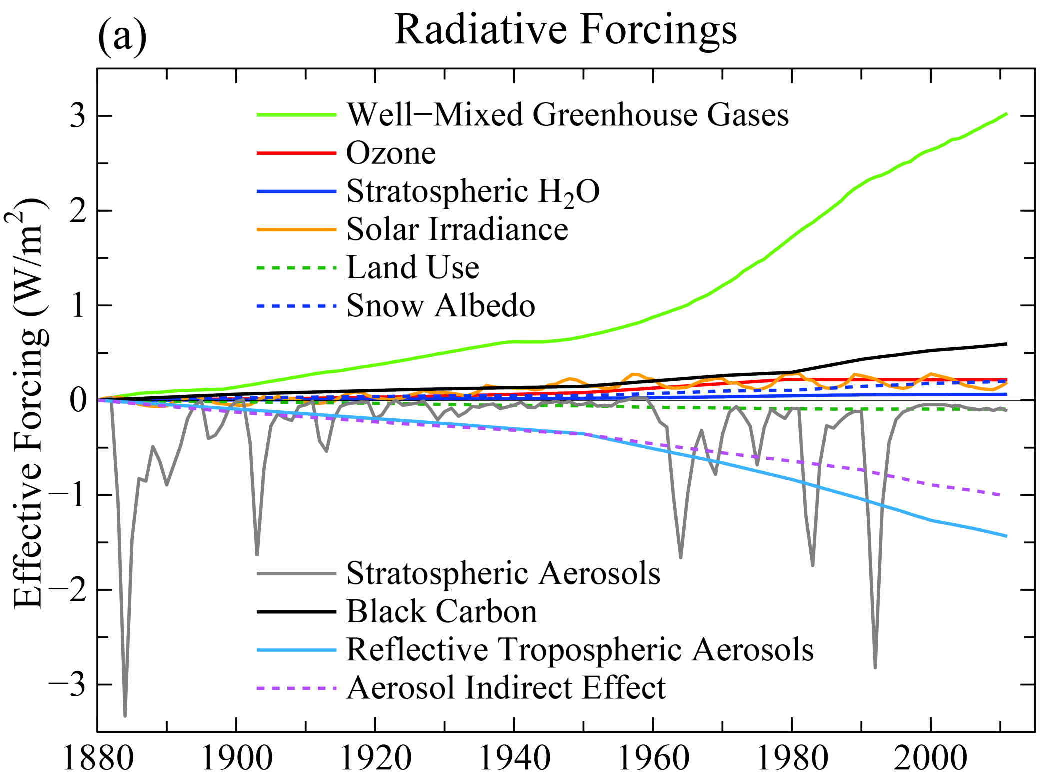


Figure 12: Radiative forcings as a function of time. From data.giss.nasa.gov.

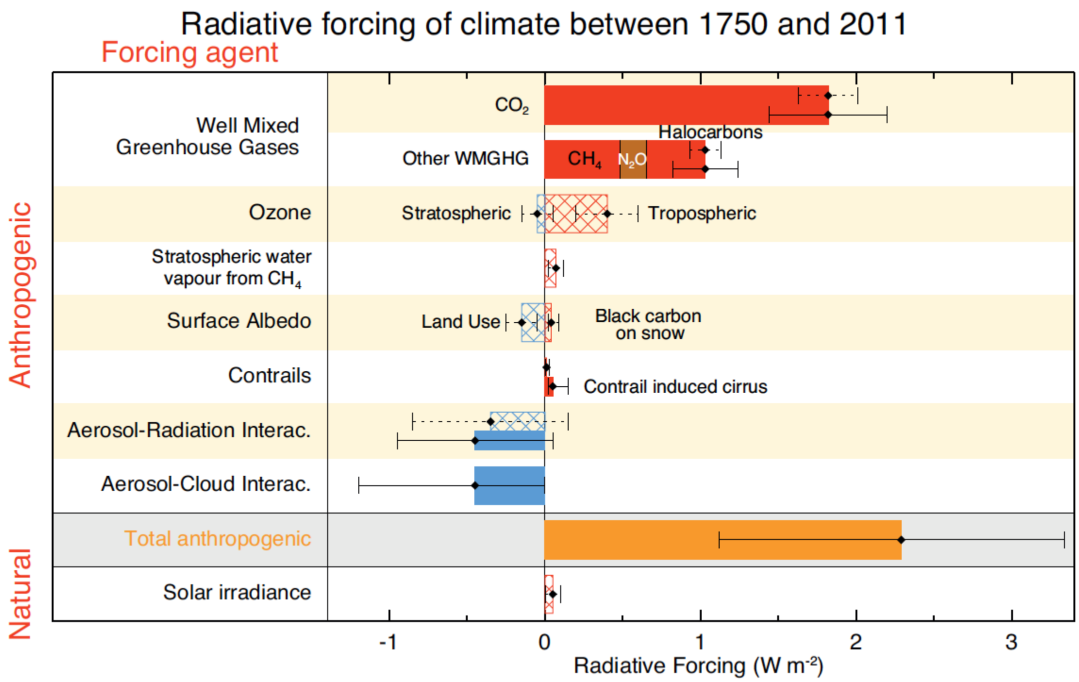
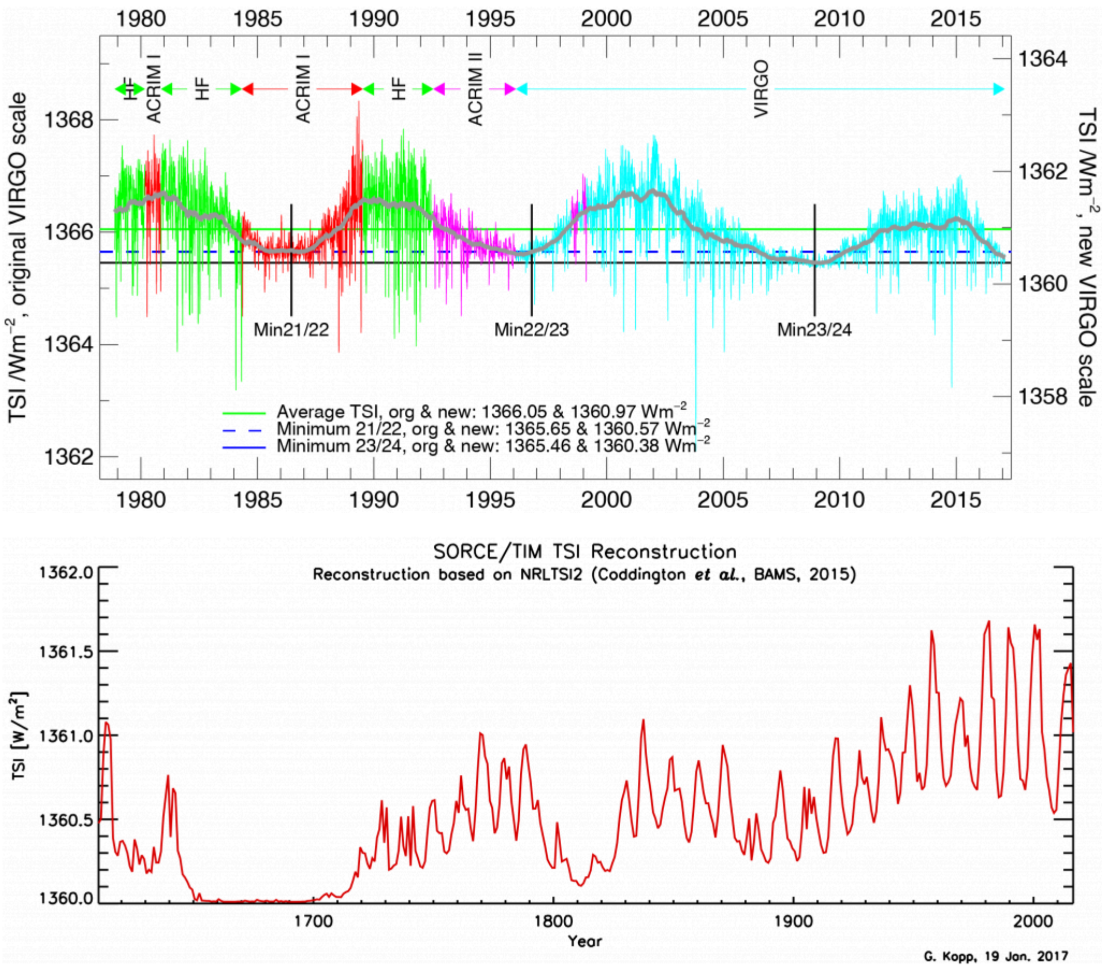


Figure 13: Summary of radiative forcings. From IPCC (2013). Hatched bars denote low confidence and scientific understanding.

Figure 14: Total solar irradiance variations.
[Top:]() measurements based on various satellites as a function of time.
[Bottom:]() longer term reconstructions.



Feedback Processes

A feedback process is a modifier of climate change. It can be defined as a process that can amplify or dampen the response to the initial radiative forcing through a feedback loop. In a feedback loop the output of a process is used to modify the input (Fig. 15). By definition, a positive feedback amplifies and a negative feedback dampens the response. In our case, the input is the radiative forcing (ΔF) and the output is the global average temperature change (ΔT).

Figure 15: Schematic illustration of fast climate feedback loops (left) and their effects on the vertical temperature distribution in the troposphere (right).

One or more interactive elements has been excluded from this version of the text. You can view them online here: https://open.oregonstate.education/climatechange/?p=104#video-104-2

Imagine talking into a microphone. A positive feedback process works like the amplifier that makes your voice louder. It can lead to a runaway effect if no or only weak negative feedback processes are present. If you hold the microphone too close to the speaker, the runaway effect can result in a loud noise. Early in Earth's history, between about 500 million and 1 billion years ago, Earth may have experienced a runaway effect into a completely ice and snow covered planet called 'Snowball Earth' caused by the ice-albedo feedback (see below). An example of a negative feedback process would be talking into a pillow. This makes your voice quieter. A negative feedback is stabilizing. It prevents a runaway effect. Both positive and negative feedback processes operate in the climate system. (Try to imagine talking into multiple pillows and microphones.) In the following we will discuss the most important ones.

Let's assume we have an initial positive forcing ($\Delta F > 0$) as illustrated in Fig. 15. As a response, temperatures in the troposphere will warm. Since the troposphere is well mixed, we can assume that the warming is uniform ($\Delta T_s = \Delta T_a > 0$). Thus, the upper troposphere will warm, which will lead to increased emitted terrestrial radiation ($\Delta ETR_a > 0$) to space. Increased heat loss opposes the forcing and leads to cooling. This is the **Planck feedback**, and it is negative. Equilibrium will be achieved if $\Delta ETR_a = \Delta F$. Since $\Delta ETR_a = ETR_{a,f} - ETR_{a,i}$ is the difference between the final $ETR_{a,f} = \sigma(T_a + \Delta T_a)^4$ and the initial $ETR_{a,i} = 240\ Wm^{-2}$ as calculated above (e.g. Fig. B2.1), we can calculate the surface temperature change due to the forcing and the Planck feedback $\Delta T_{pl} = \Delta T_a = [(ETR_{a,i} + \Delta F)/\sigma]^{1/4} - T_a$. For a doubling of atmospheric CO_2 $\Delta F = 3.7\ Wm^{-2}$ this results in $\Delta T_{pl} = 1$ K.

Thus, if only the Planck feedback was operating and everything else would be fixed, a doubling of CO_2 would result in a warming of about 1 K. However, warmer air and surface ocean temperatures will also lead to more evaporation. The amount of water vapor an air parcel can hold depends exponentially on its temperature. This relationship, which can be derived from classical thermodynamics, is called the Clausius-Clapeyron relation (Fig. 16). Since most of Earth is covered in oceans there is no lack of water available for evaporation. Therefore, it is likely that warmer air temperatures will lead to more water vapor in the atmosphere. Because water vapor is a strong greenhouse gas this will lead to an additional reduction in the amount of outgoing longwave radiation and therefore to more warming. Thus, the **water vapor feedback** is positive. If we assume again that the temperature change is uniform with height the troposphere will warm by an additional amount ΔT_{wv} due to the water vapor feedback (red line in Fig. 15).

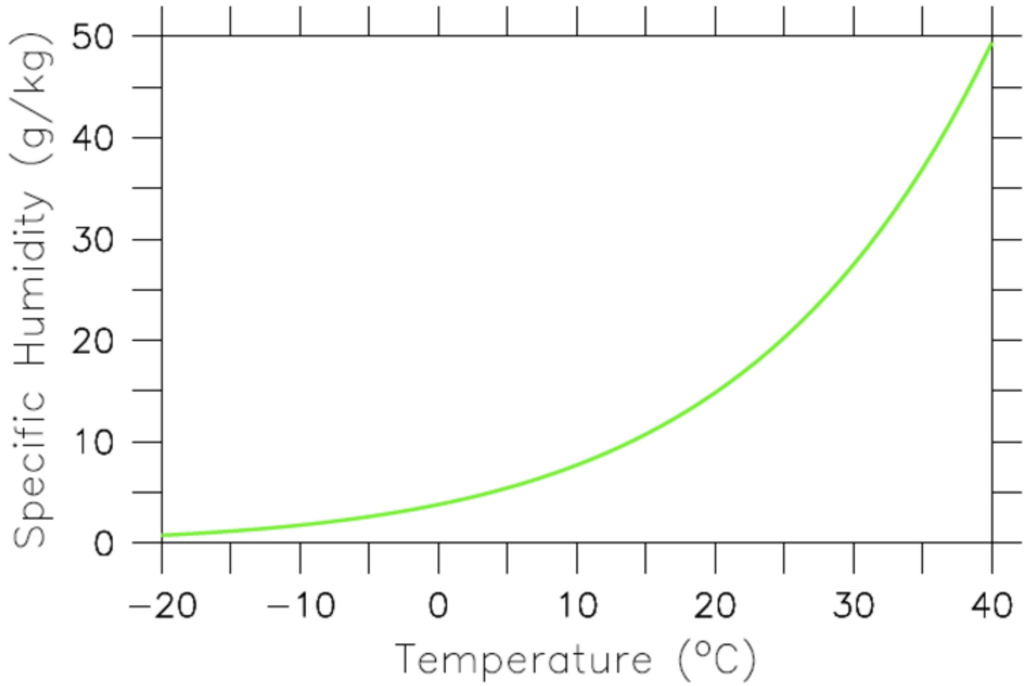


Figure 16: The Clausius-Clapeyron relation describes the amount of water vapor (in g water per kg of moist air) that air at saturation can hold as a function of temperature. The water-holding capacity of air increases approximately exponentially with temperature such that a 1 degree C warming leads to a 7% increase in humidity. All points long the green line represent 100% relative humidity. The lower a point is below the green line the lower its relative humidity will be. E.g. a point half way between the green line and the zero line would have 50% relative humidity.

Increased amounts of water vapor in the atmosphere also imply increased vertical transport of water vapor and thus more latent heat release at higher altitudes where condensation occurs. This warms the air aloft $\Delta T_{lr} > 0$. In contrast, at the surface increased evaporation leads to cooling $\Delta T_{lr} < 0$. Thus, the lapse-rate, $\Gamma = \Delta T/\Delta z$, which is the change in temperature with height in the atmosphere, is expected to decrease. Warming of the upper atmosphere will increase outgoing longwave radiation. Therefore, similarly to the Planck feedback, the **lapse rate feedback** is negative.

Since both the water vapor and the lapse rate feedback are caused by changes in the hydrologic cycle, they are coupled. This results in reduced uncertainties in climate models if the combined water vapor plus lapse rate feedback is considered rather than each feedback individually (Soden and Held, 2006).

Warming surface temperatures will also cause melting of snow and ice. This decreases the albedo and thus it increases the amount of absorbed solar radiation, which will lead to more warming. Thus, the **ice-albedo feedback** is positive. Our simple energy balance model 2 from above can be modified to include a temperature dependency of the albedo, which can exhibit a runaway transition to a snowball Earth and interesting hysteresis behavior. Hysteresis means that the state of a system does not only depend on its parameters but also on its history. Transitions between states can be rapid even if the forcing changes slowly.

Warming will also likely change clouds. However, no clearly understood mechanism has been identified so far that would make an unambiguous prediction how clouds would change in a warmer climate. Comprehensive climate models predict a large range of **cloud feedbacks**. Most of them are positive, but a negative feedback cannot be excluded at this point. The cloud feedback is the least well understood and the most uncertain element in climate models. It is also the source of the largest uncertainty for future climate projections.

Climate models can be used to quantify individual feedback parameters γ_i. They are calculated as the change in the radiative flux at the top-of-the-atmosphere ΔR_i divided by the change of the controlling variable Δx_i: $\gamma_i = \Delta R_i/\Delta x_i$. E.g. to quantify the Planck feedback the controlling variable $\Delta x_i = \Delta T$ is atmospheric temperature. The atmospheric temperature is increased everywhere by $\Delta T = 1$ K and then the radiative transfer model calculates ΔR at every grid point of the model, that is at all latitudes and longitudes, and then averages over the whole globe. This results in ΔR_{pl} and $\gamma_{pl} = \Delta R_{pl}/\Delta T$. All individual feedback parameters can be added to yield the total feedback $\gamma = \gamma_{pl} + \gamma_{wv} + \gamma_{lr} + \gamma_{ia} + \gamma_{cl}$. The total feedback has to be negative to avoid a runaway effect. The strongest, most precisely known feedback is the Planck feedback, which is about $\gamma_{pl} = -3.2$ Wm^{-2}K^{-1}. Estimates for the other feedbacks are about $\gamma_{wv} + \gamma_{lr} \cong +1$ Wm^{-2}K^{-1} for the combined water vapor / lapse rate feedbacks, $\gamma_{ia} = +0.3$ Wm^{-2}K^{-1} for ice-albedo feedback, and $\gamma_{cl} = +0.8$ Wm^{-2}K^{-1} for the cloud feedback. This gives for the total feedback γ values from about −0.8 to about −1.6 Wm^{-2}. The total feedback parameter can be used to calculate the climate sensitivity.

Climate Sensitivity

The climate sensitivity $\Delta T_{2\times}$ is usually defined as the global surface temperature change for a doubling of atmospheric CO_2 at equilibrium and it includes all fast feedbacks discussed above. Current best estimates are $\Delta T_{2\times} \cong 3$ K, however it ranges from about 1.5 to about 4.5 K. This large uncertainty is mostly due to the large uncertainty of the cloud feedback. Sometimes the climate sensitivity $S_C = -1/\gamma$ is reported in units of K/(Wm^{-2}). Since we know the forcing for a doubling of CO_2 $\Delta F_{2\times} = 3.7$ Wm^{-2} quite well, one can be calculated from the other using $S_C = \Delta T_{2\times}/\Delta F_{2\times}$. *For* $\Delta T_{2\times} \cong 3$ K, $S_C \cong 0.8$ K/(Wm^{-2}), *and* $\gamma \cong 1.2$ Wm^{-2}K^{-1}. Note that slow feedbacks associated with the growth and melting of ice sheets or changes in the carbon cycle are not included in these numbers. Since we would expect those feedbacks to be also positive we can expect an even higher climate sensitivity for longer timescales (hundredths to thousandths of years).

Our definitions of radiative forcings and feedbacks above are not clear cut. They are based on existing climate models and the processes included in them. For example changes in atmospheric CO_2 concentrations over long paleoclimate timescales can be thought of as a feedback rather than a forcing since the ultimate forcing of the ice age cycles are changes in the Earth's orbital parameters and hence the seasonal distribution of solar radiation.

Questions

- What is a budget equation? Describe it using your own words.
- When is a budget in balance?
- What is electromagnetic radiation?
- What is a photon?
- Which has shorter and which longer wavelengths: ultraviolet, visible, or infrared radiation?
- Which of the previous three wavelengths photons is highest/lowest in energy?
- How does electromagnetic radiation interact with matter?
- What is a blackbody?
- What are the differences between the blackbody radiation curves for the sun and Earth?
- What is the Stefan-Boltzmann law? Write down the formula and explain the variables.
- What is the 'bare rock' model of the Earth's energy balance and why did it fail to correctly predict Earth's surface temperature?
- What is temperature? Describe it using your own words.
- What is the greenhouse effect?
- What are greenhouse gases?
- What are Earth's two most important greenhouse gases?
- What is the total solar irradiance (TSI) at the top of the atmosphere on a surface perpendicular to the sun's rays (in W/m^2)?
- What is the incoming solar irradiance (S) at the top of the atmosphere averaged over the whole Earth and over one year (in W/m^2)?
- What percentage of the incident solar radiation at the top-of-the-atmosphere is absorbed by the atmosphere?
- What percentage of the upward terrestrial radiation emitted from the surface is absorbed by the atmosphere?

Videos

Lecture: Radiation
Lecture: Greenhouse Effect
Lecture: Energy Budget
Lecture: Forcings
Lecture: Feedbacks
Lecture: Climate Sensitivity
Lab: Simple Climate Model

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5. Carbon

Due to the importance of CO_2 as a greenhouse gas the carbon cycle is a crucial part of the climate system. Since carbon exchanges with the biosphere, biological processes need to be considered in climate science. The carbon cycle is part of the broader biogeochemical cycles, which include other biologically important chemical elements such as nitrogen and oxygen.

a) The Natural Carbon Cycle

Carbon exchanges relatively rapidly between three large reservoirs: the ocean, the atmosphere, and the land (Fig. 1). Of those the **ocean** contains the most carbon: almost 40,000 Pg. Most of the carbon in the ocean is in the form of Dissolved Inorganic Carbon (DIC), and most DIC resides in the intermediate and deep layers because those depths make up most of its volume. Marine biota are important in transferring carbon from the surface to the deep ocean, but their biomass is very small because they consist mainly of microscopic algae called phytoplankton. Phytoplankton build the base of the ocean's food web through photosynthesis. They have adapted to be tiny and light, so as not to sink to the sea floor. They need to stay near the sunlit surface to photosynthesize. Below about 100 m depth light levels get too low due to **absorption** by sea water. The deep ocean is therefore dark but organic matter sinks there from the surface in various forms, e.g. as fecal pellets of zooplankton. Below the surface, the sinking dead organic matter is respired by bacteria and returned into the inorganic carbon pool. This is called the ***biological pump*** because it removes carbon from the surface and atmosphere and sequesters it in the deep ocean, where it can stay for hundreds to thousands of years. Dissolved CO_2 gas in sea water is part of the DIC pool. It exchanges with the atmosphere about 80 Pg of carbon per year. Ocean-atmosphere gas exchange depends on the difference between surface ocean and atmospheric partial pressures (pCO_2; in this book we use pCO_2 and CO_2 concentration synonymously in units of parts per million or ppm) and therefore leads to a strong and relatively rapid coupling of the atmospheric CO_2 concentration to the surface ocean.

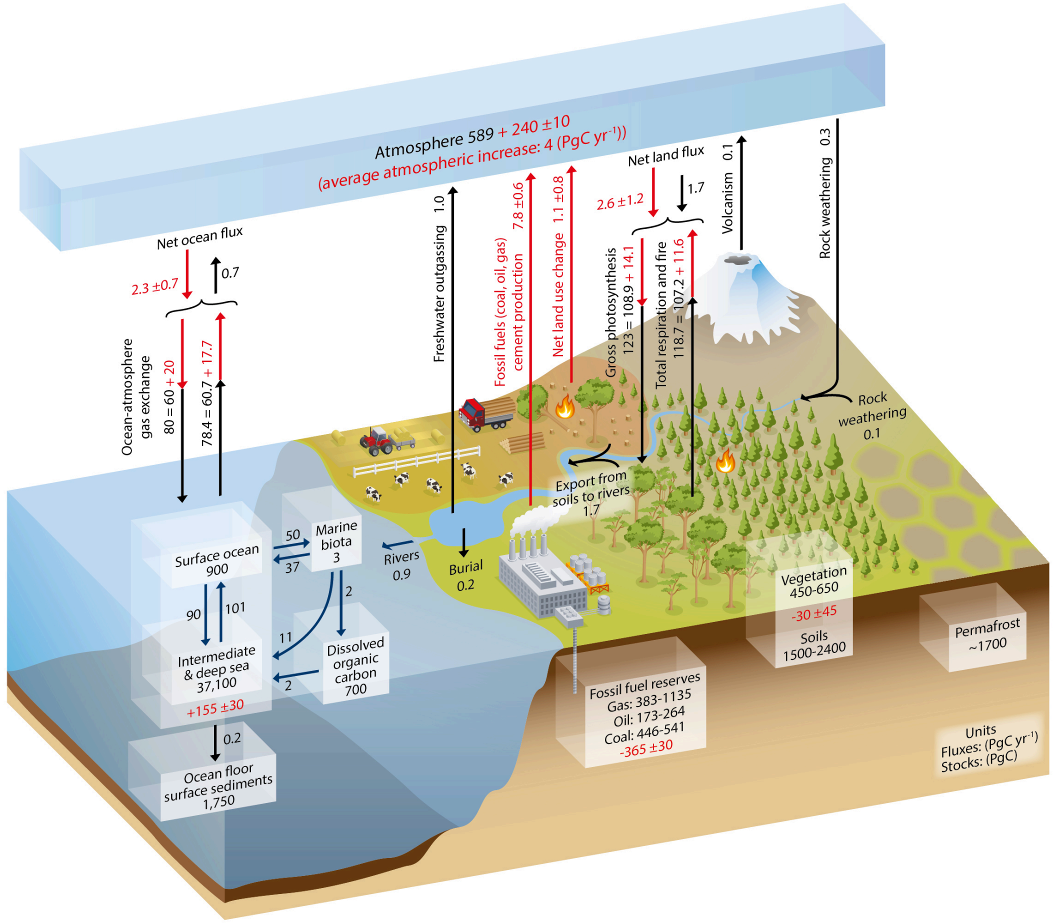


Figure 1: The Carbon Cycle. Numbers in boxes represent reservoir mass, also called carbon stocks, inventories, or storage in PgC (1 Pg = 10^{15}g = 1 Gt). 2.1 PgC = 1 ppm atmospheric CO_2. Numbers next to arrows indicate fluxes in PgC/yr. Black numbers and arrows represent estimates of the natural (pre-industrial) carbon cycle. Red numbers and arrows indicate estimates of anthropogenic effects for 2000-2009. From Ciais et al. (2013). [Image from ipcc.ch](). See additional material at the end of this chapter for a figure showing the sizes of the reservoirs and fluxes in proportion to the size of the boxes and arrows.

The second biggest of the three rapidly exchanging carbon reservoirs is the **land**, which contains about 4,000 Pg of carbon. On land carbon is stored in living vegetation, in soils, and in permafrost. Since land plants don't have the problem of sinking out of the light they can grow big and contain large amounts of carbon, such as trees. Therefore, much more carbon is stored in living biomass on land (~500 Pg) than in the ocean (~3 Pg). However, even more carbon is stored in soils and permafrost.

More than 100 Pg of carbon are removed from the atmosphere each year by photosynthesis of land plants and turned into organic matter. Organic matter cycles through the land food web and eventually gets into the soil carbon pool where it decomposes. Like in the ocean, bacteria and heterotrophic organisms on land respire organic carbon and turn it back into inorganic CO_2. Land carbon uptake and release does not depend strongly on atmospheric CO_2 concentrations. They depend more on water availability and temperature, respectively. Plant growth on land is strongly water limited and respiration rates strongly depend on temperature. However, CO_2 increases the water use efficiency of land plants because at higher CO_2 concentrations they don't need to open the stomata as much as at lower CO_2 concentrations. Stomata are small openings in the cells that allow CO_2 to enter, but they also allow water to leave in a process called transpiration (see figure in box below). Thus, at higher CO_2 levels plants can grow more for the same amount of water usage.

The atmospheric carbon reservoir is relatively small compared to the ocean, which is ~40 times bigger, and the land, which is ~10 times bigger. However, the **atmosphere** is crucial in linking land and ocean through rapid exchanges.

Box 1: Photosynthesis and Respiration

Photosynthesis (Fig. B1.1) is the process by which autotrophic organisms (plants, algae, and many bacteria) produce organic matter and oxygen from CO_2 and water using light as an energy source.

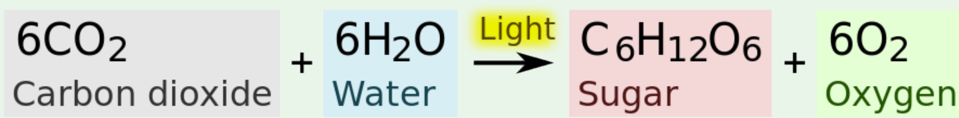


Figure B1.1: Chemical Reaction for Photosynthesis. Carbon dioxide and water is converted to carbohydrates and oxygen. This process uses energy.

Respiration (Fig. B1.2) is the reverse process by which heterotrophic organisms (bacteria, fungi, animals, and humans) oxidize organic carbohydrates to derive their energy resulting in CO_2 and water.

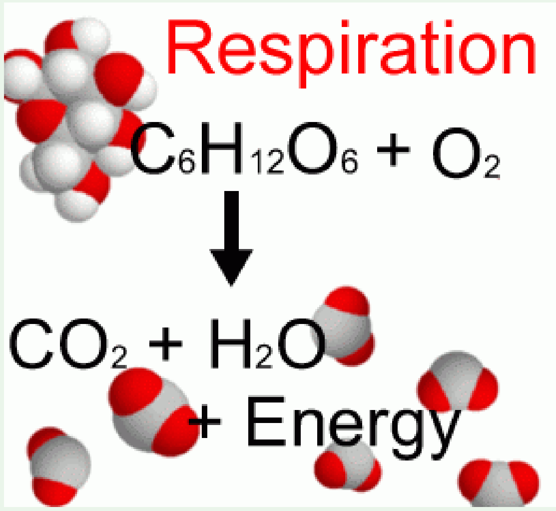


Figure B1.2:Chemical Reaction for Respiration. Carbohydrates and oxygen are converted to CO_2 and water. This process releases energy.

In order to photosynthesize, land plants have to take up CO_2 from the air. They do this by opening little pores called stomata, through which not only CO_2 can enter, but also water and oxygen can leave the cell (Fig. B1.3).

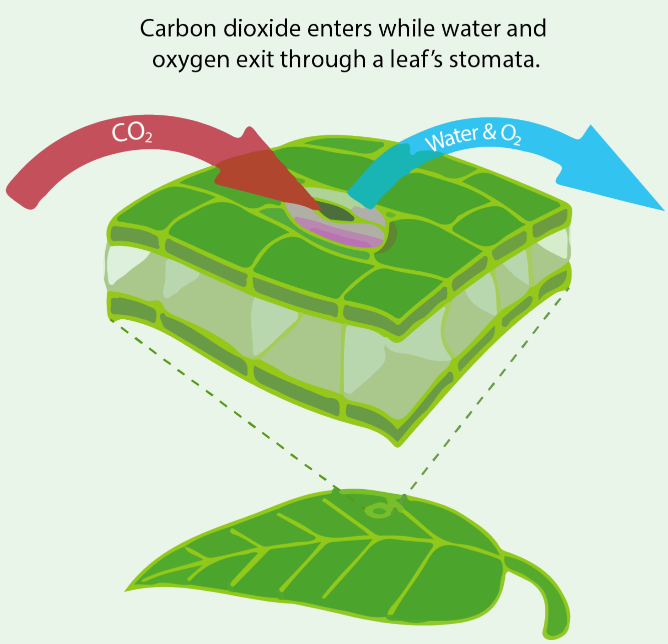


Figure B1.3: Illustration of Gas Exchange Through Stomata in a Leaf.

b) Anthropogenic Carbon

Human effects on the global carbon cycle have been relatively limited before the industrial revolution, although some emissions from land-use change such as deforestation may have been going on for hundreds or thousands of years. During the last 100 years or so, however, rapid burning of fossil fuels such as coal, gas, and oil have caused a massive perturbation (Fig. 2). This perturbation is perhaps most evident in the atmosphere where CO_2 concentrations have increased by more than 40 %. In chapter 3 we have seen that current levels of atmospheric CO_2 have been unprecedented for the last 800,000 years, but reconstructions going back further in time indicate that the last time Earth's atmosphere had about 400 ppm CO_2 is about 3 million years ago.

Currently humans emit about 10 billion tons of carbon into the atmosphere per year mostly from fossil fuel burning (~90 %). However, deforestation continues to be a significant contribution (~10 %). Anthropogenic carbon emissions from fossil fuel burning have increased rapidly after World War Two. The ocean has taken up about 40 % (155/395) of all anthropogenic carbon emissions hitherto. 60 % (240/395) have stayed in the atmosphere, whereas the land comes out in a wash because of loss due to deforestation and gain due to recent regrowth.

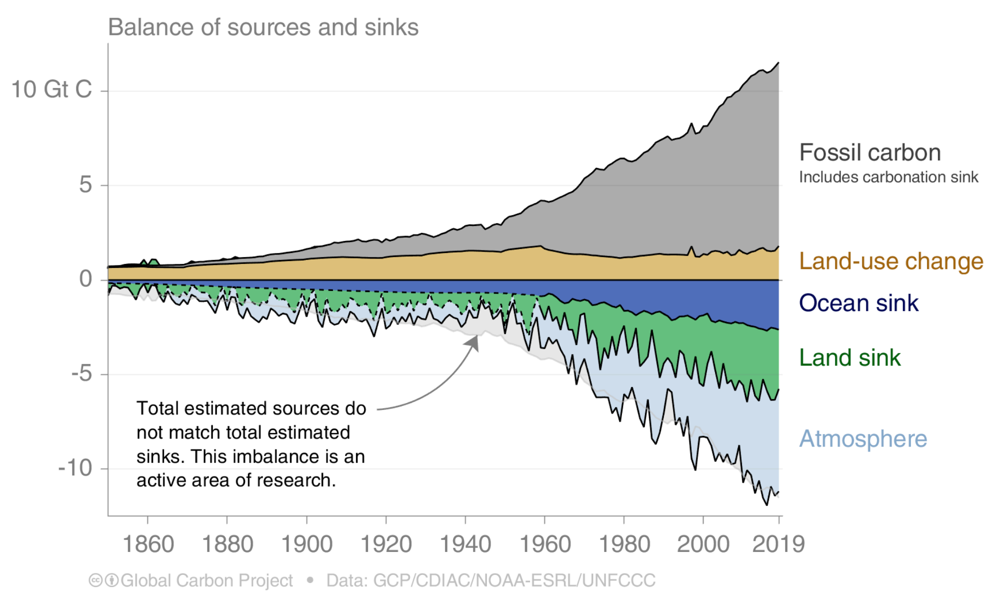


Figure 2: Anthropogenic Carbon Sources and Sinks in Billion Metric Tons of Carbon Per Year (Gt C/yr) as a function of the year. Positive numbers show sources from fossil fuel burning and cement production (gray) and deforestation (yellow). Negative numbers show uptake by land (green), atmosphere (light blue), and ocean (dark blue). From Friedlingstein et al. (2019). Image from globalcarbonproject.org.

Anthropogenic carbon emissions have increased rapidly during the early 2000's but in recent years they have flattened out (Fig. 3) mostly because of emissions from China showed a similar behavior

(Fig. 4). About half of all carbon put into the atmosphere by humans since the industrial revolution (cumulative emissions) was done so in the last 30 years. Cumulative emissions are the grey and brown areas in Fig. 2. Together they amount to about 500 GtC or half a trillion metric tons. As we will see later cumulative carbon emissions determine the global temperature increase.

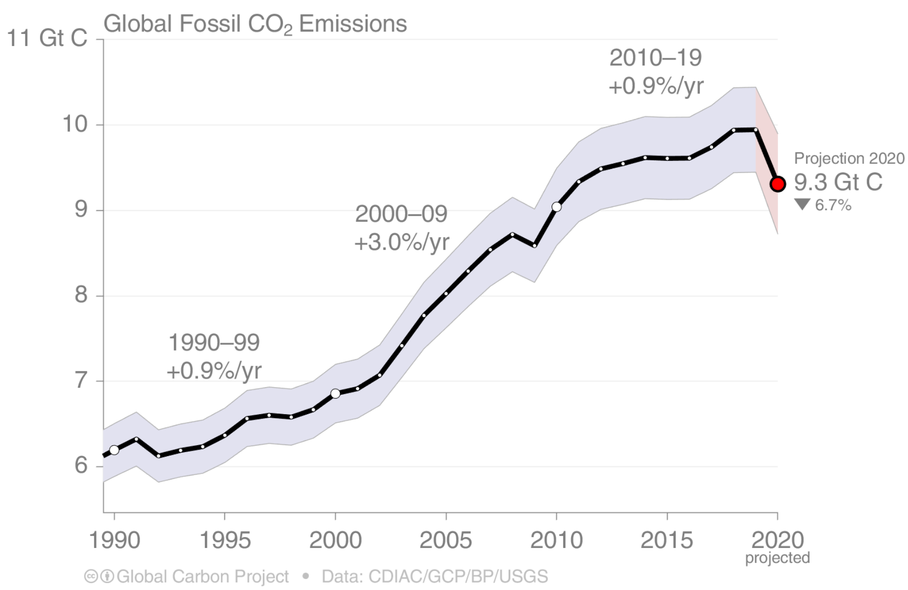


Figure 3: Estimates of carbon emissions from fossil fuel burning and cement production since 1990. From globalcarbonproject.org.

The effects of the financial and economic crisis is seen in the dip in global carbon emissions in 2009 caused by emission reductions in the US and Europe (Fig. 4), whereas emissions continued to increase in China until 2013 after which they stayed constant. A similar dip can be expected for the current economic slowdown due to the coronavirus pandemic.

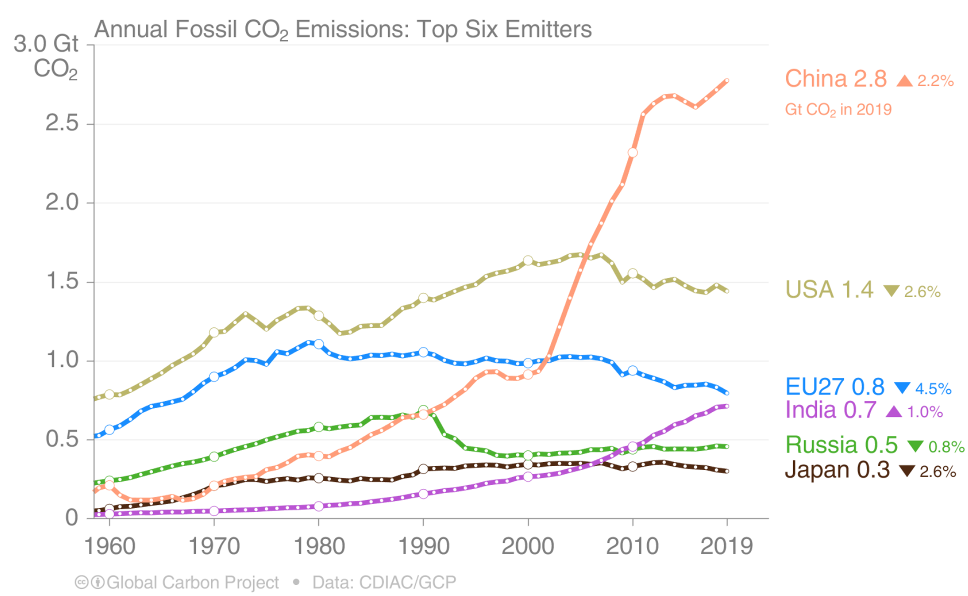


Figure 4: Carbon emissions in GtC/yr (units in the figure labels are incorrect) from the top four fossil fuel emitters. From globalcarbonproject.org.

Human caused carbon emissions are mainly from burning of fossil fuels, whereas cement production contributes only about 6 % (Fig. 5). Burning of coal, oil, and gas have all increased substantially during the last 50 years. The increase in emissions from China during the first decade of the 21st century was fueled mainly by coal burning.

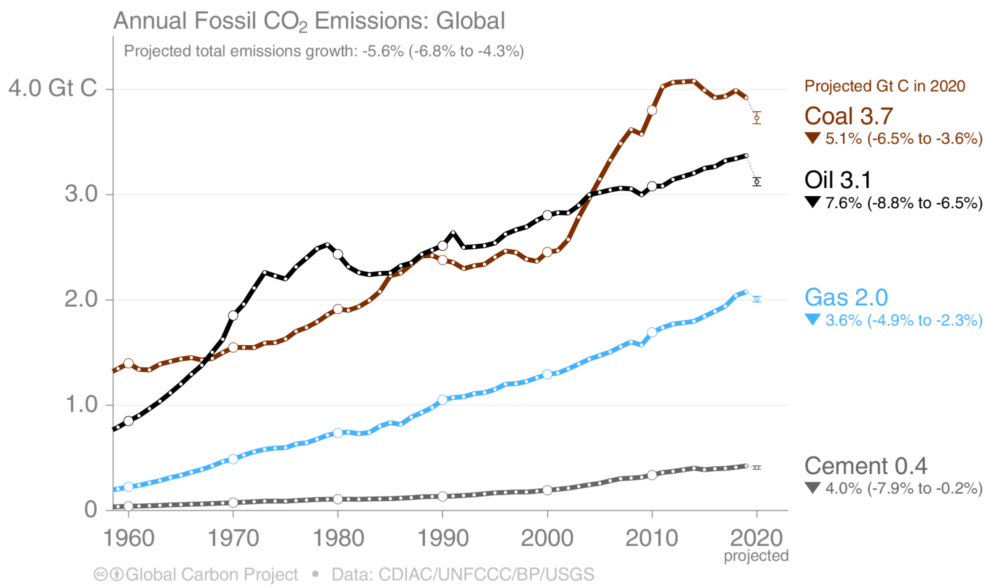


Figure 5: Contributions to fossil fuel emissions in GtC/yr from coal, oil, and gas burning and cement production. From globalcarbonproject.org.

Among the four top emitters the US is the one with the largest emissions per person (Fig. 6). The average US American emits more than 4 metric tons of carbon into the air each year. This is more than twice the emissions per person in Europe or China, more than three times the average emissions world-wide, and about ten times the emissions from a person in India. The US is responsible for 25% of all carbon emitted in the past (cumulative emissions) although it makes up only 4% of the world's population. Europe, which accounts for about 10% or the world's population, has emitted more than 22% of all carbon. This figure shows other countries.

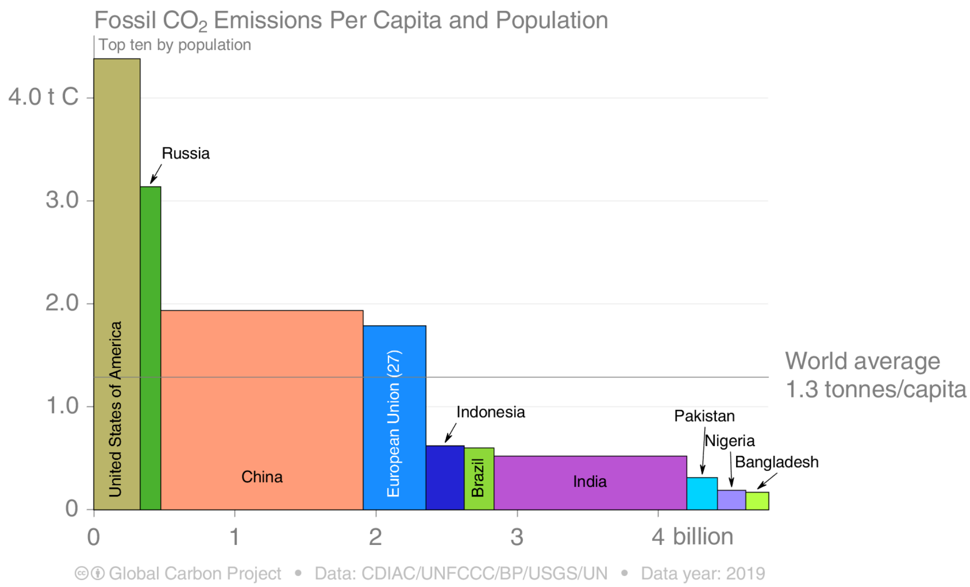


Figure 6: Carbon emissions per person in tC/yr. The width of the boxes represents the population. From globalcarbonproject.org.

How do we know that the rising CO_2 concentrations in the atmosphere are from human activities? There are several independent lines of evidence. The first comes from economic data. Since fossil fuels are traded internationally we know how much oil, coal, and gas a country imports and uses. The data shown in Figs. (2) through (6) are based on these estimates. The shaded area in Fig. (3) indicates the error bars of those estimates. Not all countries publish and make their data available, which leads to these uncertainties. However, they are relatively small such that the emissions are known to within about a 5 % error margin.

Box 2: Carbon Isotopes

Carbon exists as three isotopes. The most common carbon-12 (^{12}C) with 6 protons and 6 neutrons, the rarer carbon-13 (^{13}C) with an additional neutron, and carbon-14 (^{14}C) or radiocarbon with two additional neutrons. ^{14}C is radioactive and decays with a half-life of 5,730 years.

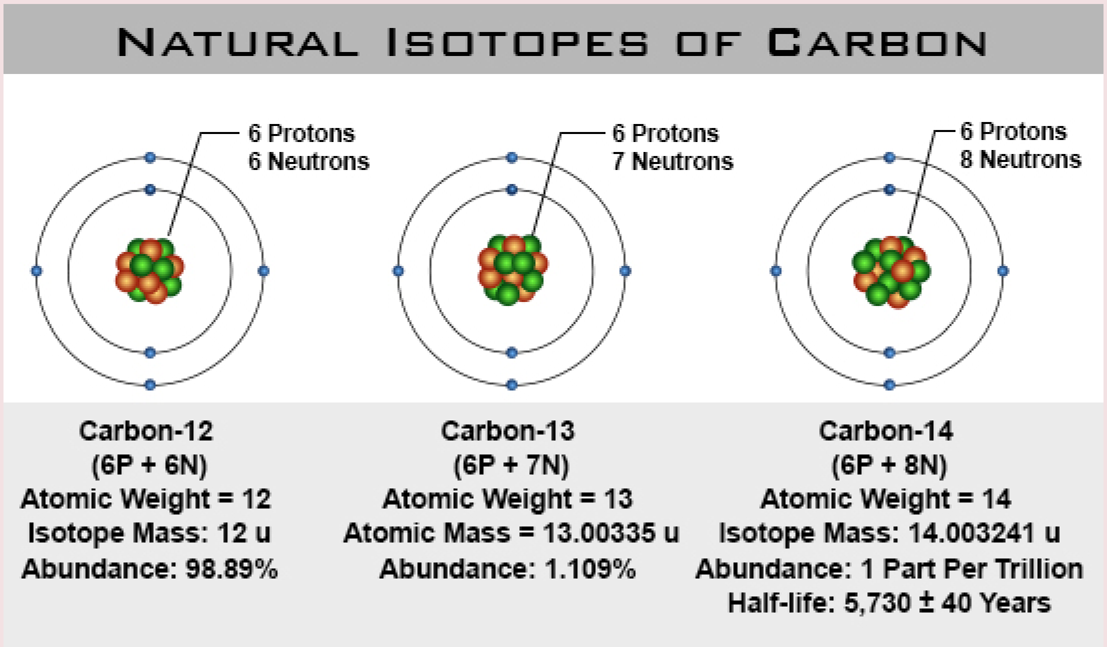


Figure B2.1: Illustration of the Atomic Structure of Carbon Isotopes. The nucleus consists of protons (red) and neutrons (green). It is surrounded by electrons (blue).

Plants and algae fractionate carbon isotopes by about 20 ‰ during photosynthesis such that they preferentially take up the light ^{12}C. Pre-industrial $\delta^{13}C$ values of atmospheric CO_2 were about -6.5 ‰. Thus plant and soil carbon have $\delta^{13}C$ values of around -27 ‰.

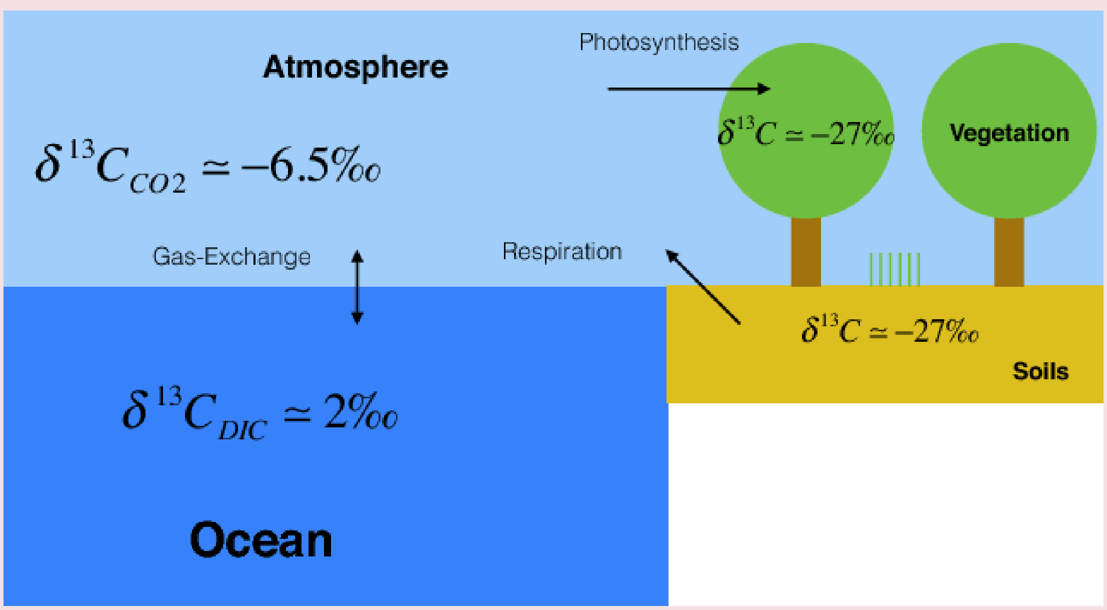
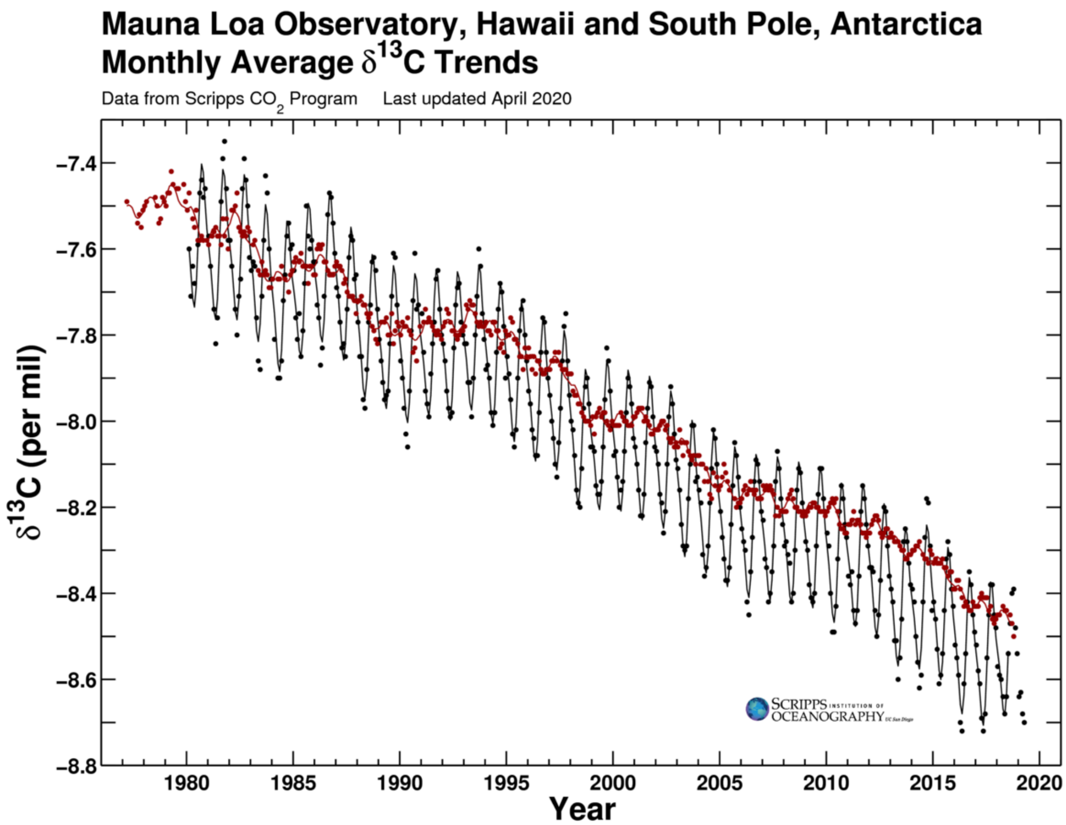


Figure B2.2: Schematic Distribution of Carbon Isotopes in the Carbon Cycle Components.

The delta notation is analogous to that of oxygen isotopes discussed in chapter 3. $\delta^{13}C = R/R_{std} - 1$, where $R = {}^{13}C/{}^{12}C$ is the ratio of the heavy over the light isotope and R_{std} is that of a standard.

The second line of evidence is based on carbon isotope measurements. Fractionation during photosynthesis leads to plants and algae having very depleted $\delta^{13}C$ values (see box Carbon Isotopes). Since fossil fuels are derived from ancient plants they are depleted in ^{13}C isotopes as well. Thus the addition of carbon with a very depleted ^{13}C signature to the atmosphere leads to a decrease in $\delta^{13}C$ values of atmospheric CO_2. This is observed in measurements both from ambient air and air extracted from ice cores (Fig. 7).



Mauna Loa Observatory, Hawaii and South Pole, Antarctica
Monthly Average δ13C Trends
Data from Scripps CO2 Program
Last updated April 2020
δ13C (per mil)
−7.4
−7.6
−7.8
−8.0
−8.2
−8.4
−8.6
−8.8
1980
1985
1990
1995
2000
2005
2010
2015
2020
Year
SCRIPPS INSTITUTION OF OCEANOGRAPHY

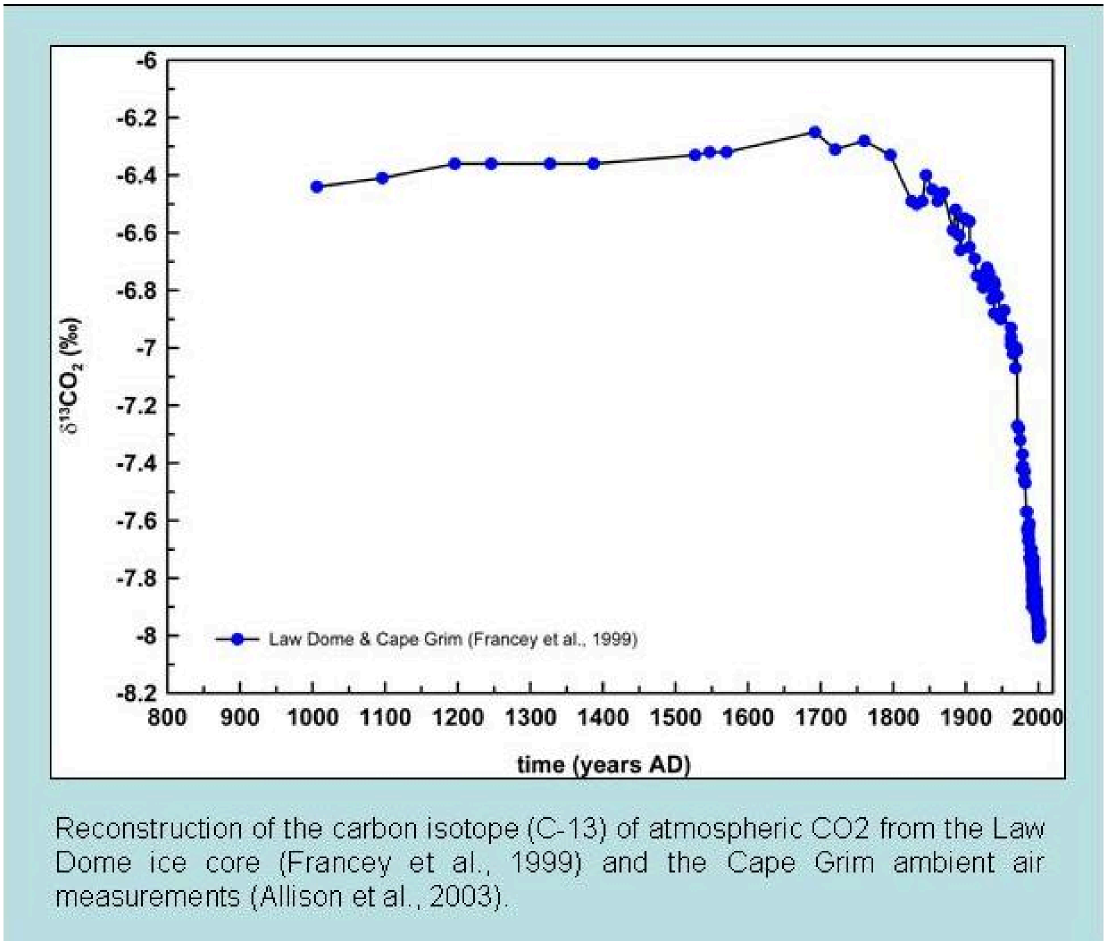


Figure 7: Carbon isotope measurements of atmospheric CO_2. Top: Ambient air data from Mauna Loa (black) and the South Pole (red). Bottom: ice core measurements.

The third line of evidence is based on oxygen measurements in air. Burning of fossil fuels has a similar chemical reaction equation that that of respiration. Carbohydrates react with oxygen to form CO_2 and water. Energy is released during this reaction. Thus, burning of carbohydrates consumes oxygen. By measuring the oxygen to nitrogen ratio in air the changes in atmospheric oxygen concentration can be detected even though they are small compared to the absolute oxygen concentrations (Fig. 8). These measurements are evidence of a massive combustion process happening on Earth right now.

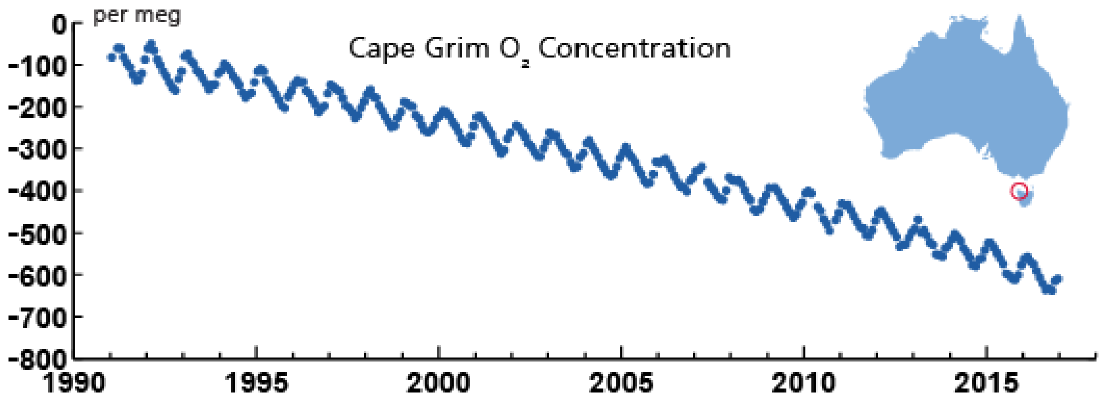


Figure 8: Measurements of atmospheric oxygen to nitrogen ratios from Cape Grim, Tasmania. Similar measurements exist from other locations. Image from scrippso2.ucsd.edu.

We conclude that humans have caused a large perturbation to the natural carbon cycle mostly by the burning of fossil fuels, which has increased atmospheric CO_2 concentrations from 280 ppm to more than 400 ppm, to levels unprecedented in Earth's history for about 3 million years. About 40 % of the anthropogenic carbon emitted so far has been taken up by the ocean, thus reducing the accumulation of CO_2 in the atmosphere.

Box 3: Residence Time

The residence time τ of a substance in a reservoir is the time required to completely replace the reservoir with its input: $\tau = X/I = X/O$ (at equilibrium I = O), where X is the reservoir size (a.k.a. stock, amount, or inventory) and I (O) is the input (output) flux. See Budget Equation Box in Chapter 4.

Exercise: Use Fig. 1 to calculate

- for the pre-industrial period the residence times of carbon (tip: sum up all the inputs or outputs to calculate I or O) in the
    - atmosphere,
    - ocean,
    - land,
    - combined ocean-atmosphere-land system, and
- the residence time of anthropogenic carbon in the combined ocean-atmosphere-land system.

c) Carbonate Chemistry and Ocean Acidification

CO_2 enters the ocean from the atmosphere through gas exchange if the partial pressure in the atmosphere is larger than the partial pressure in the ocean. It dissolves as CO_2 gas in water just like it is dissolved in your soda drink. If you look at an unopened bottle of soda, you do not see bubbles. The CO_2 molecules are emerged within a vast number of water molecules. The drink was bottled under pressure or under cold temperatures. Solubility of CO_2 like that of other gases such as oxygen depends on temperature. More gas can be dissolved in colder water (Fig. 9). This is the reason why as you warm up a soda drink it will lose CO_2. Because it is not liquid, CO_2 does not 'evaporate' into the air, but it outgasses. Evaporation implies a phase change, which does not occur in this case.

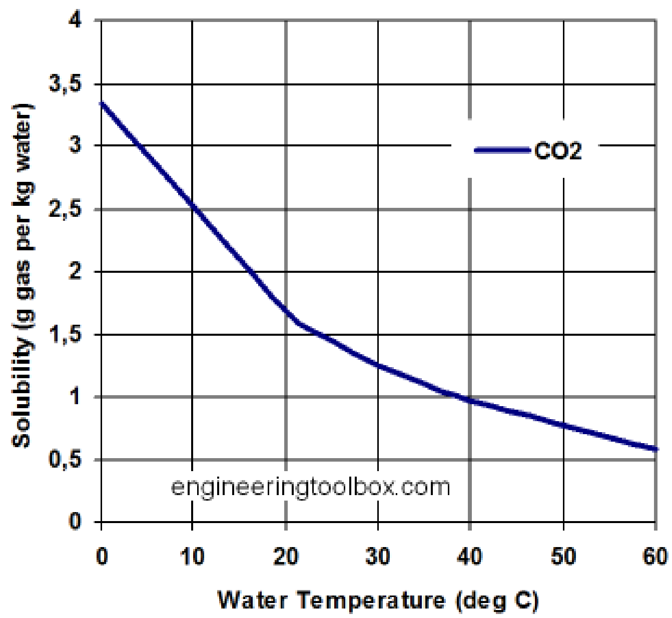


Figure 9: Solubility of CO2 in water. From engineeringtoolbox.com.

In the ocean CO_2 reacts with seawater to form carbonic acid (H_2CO_3), which dissociates into bicarbonate (HCO_3^-) and carbonate (CO_3^{2-}) ions:

(1) $$CO_2 + H_2O \Longleftrightarrow H_2CO_3 \Longleftrightarrow HCO^{3-} + H^+ \Longleftrightarrow CO_3^{2-} + 2H^+.$$

The sum of these three carbon species is called dissolved inorganic carbon (DIC = CO_2 + HCO_3^- + CO_3^{2-}) or total carbon. The equilibrium between the species depends on the pH. In the current ocean, pH is about 8.1, which leads to about 86.5 % of DIC being in the form of bicarbonate ions, 13.0 % in the form of carbonate ions, and only 0.5 % in the form of aqueous CO_2 (Fig. 10; Zeebe and Wolf-Gladrow, 2001).

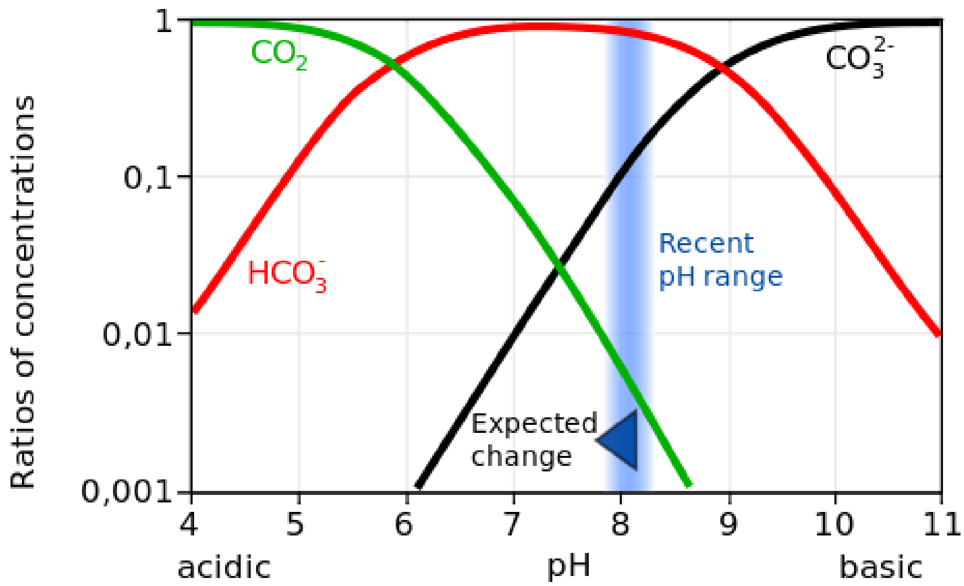


Figure 10: Ratios of carbonate species concentrations as a function of the pH. Currently average seawater has a pH of about 8.1. Therefore most carbon in the ocean is in the form of bicarbonate. Addition of anthropogenic CO_2 decreases the pH. From wikipedia.org.

Dissociation of carbonic acid into bicarbonate (baking soda) produces a hydrogen ion H^+, which decreases the pH of the water. Most hydrogen ions, however, re-combine with carbonate ions to form additional bicarbonate ions. Nevertheless, adding CO_2 to seawater increases its hydrogen ion concentration (decreases its pH) and decreases the carbonate ion concentration. This process is called ocean acidification.

Observations show that the partial pressure of surface ocean water follows closely the trend in atmospheric CO_2 (Fig. 11) indicating the uptake of anthropogenic carbon. The measurements also demonstrate that the ocean's pH decreases. Data from near Hawaii show that the pH has decreased by about 0.05 units from 1988 to 2011. Global estimates suggest a decrease by 0.2 units from preindustrial times. This corresponds to a ~30 % increase in hydrogen ions.

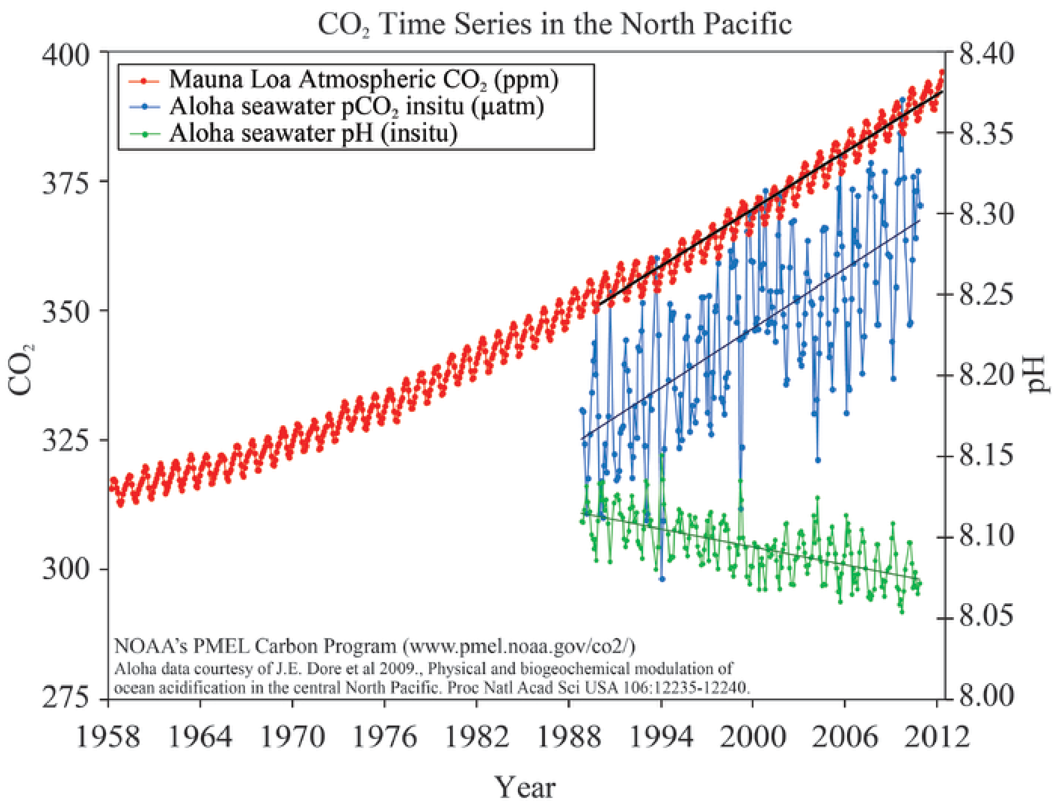


Figure 11: Measurements of seawater pCO_2 (blue) and pH (green) at ocean station Aloha near Hawaii together with the Mauna Loa record of atmospheric CO_2 (red). From pmel.noaa.gov.

The penetration of anthropogenic carbon into the ocean is largest in the North Atlantic, at mid-latitudes in the Southern Ocean and in the subtropical North Pacific (Fig. 12). As we will see below these are regions of convergence, subduction, or deep water formation.

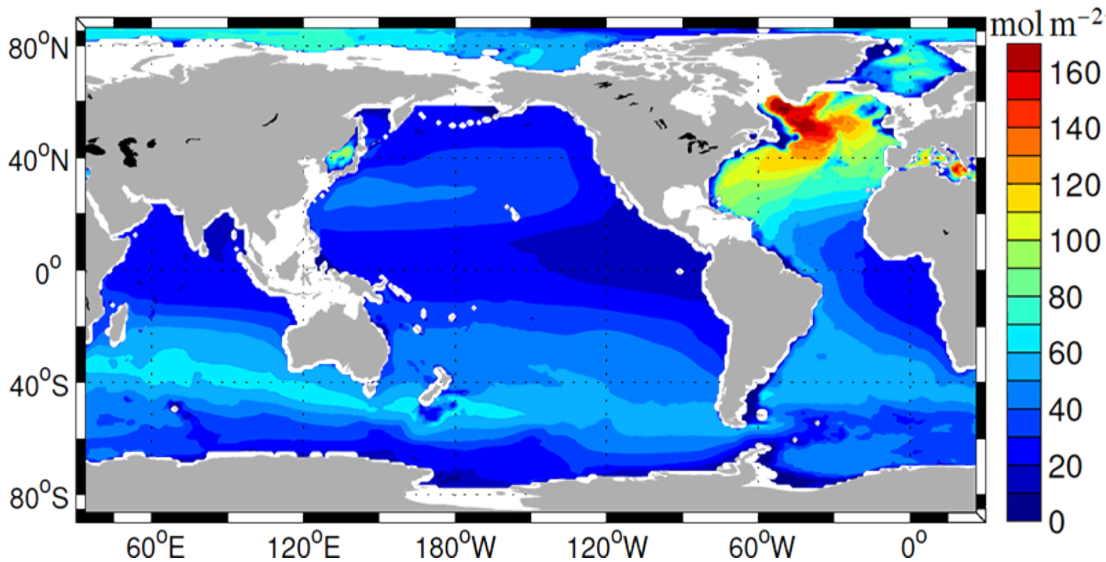


Figure 12: Anthropogenic carbon in the ocean estimated from observations (Khatiwala et al., 2013). From rapid.ac.uk.

Anthropogenic CO_2 enters the ocean at the surface. Therefore most anthropogenic carbon is in the surface layers (Fig. 13). However, measurable amounts have penetrated most of the upper kilometer of the ocean. In some regions such as the North Atlantic and in the Southern Ocean anthropogenic carbon has entered levels below 2 km depth. These are regions in the ocean where surface waters sink to great depths taking anthropogenic carbon with them.

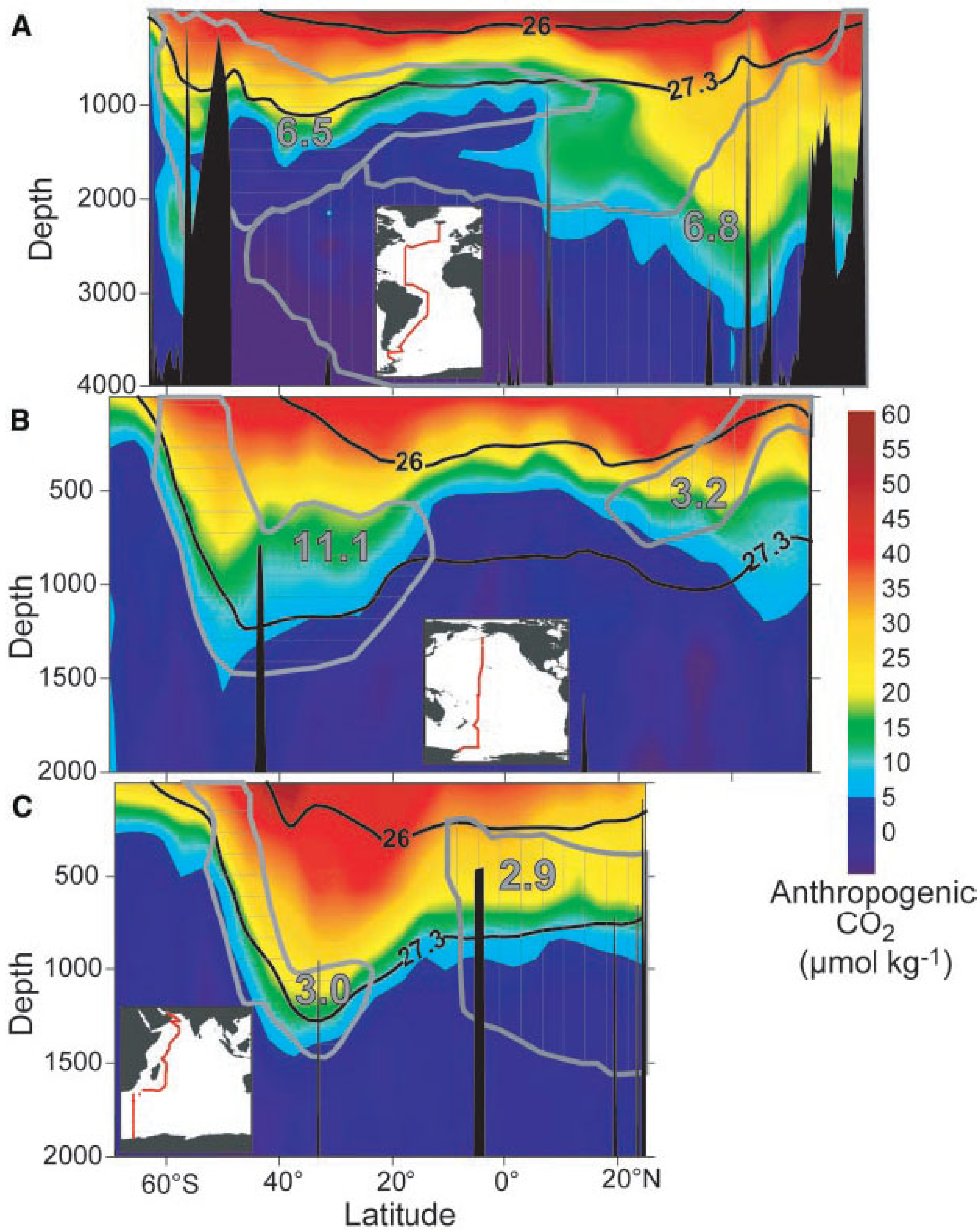
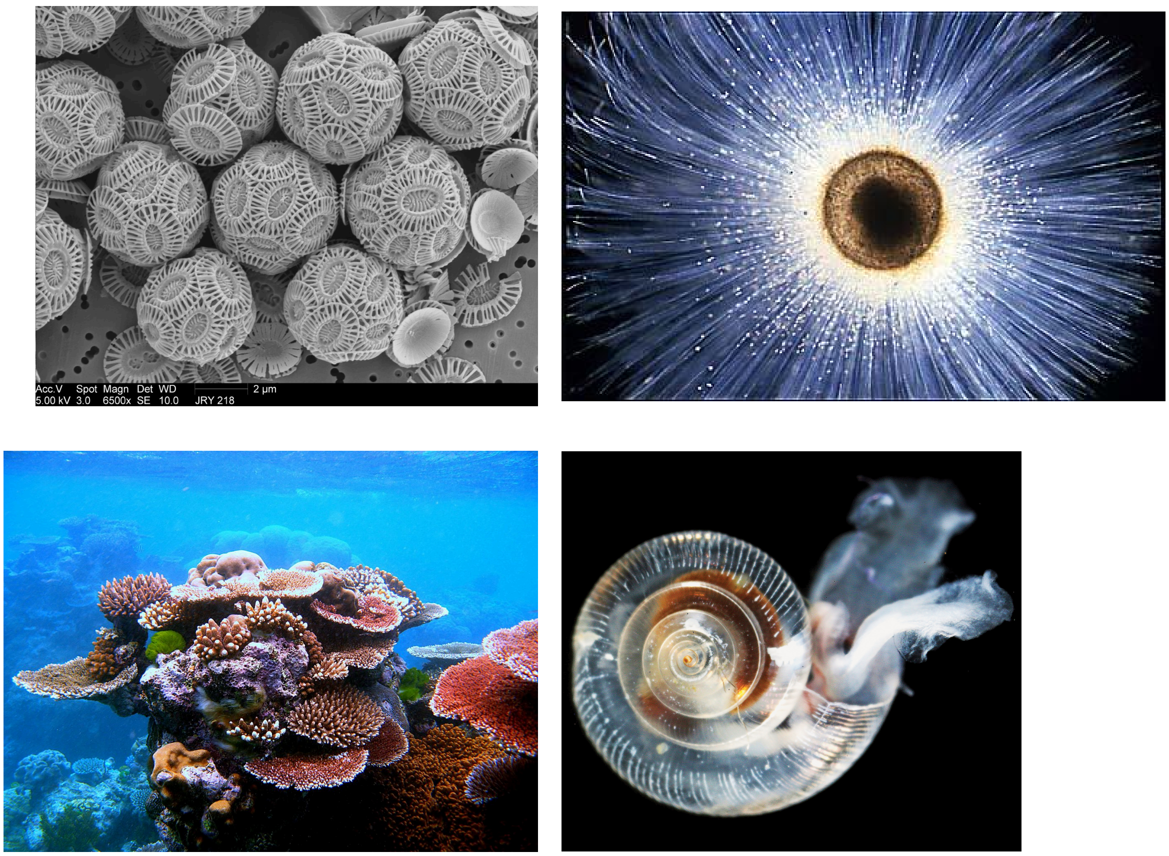


Figure 13: Latitude-depth sections of anthropogenic carbon in the Atlantic (A), Pacific (B), and Indian (C) oceans. Note that the depth scale is different in panel (A) from the other panels. Black lines denote two surfaces of constant potential density (expressed in σ_θ = density − 1000 kg m^{-3}). Movements in the ocean are preferentially along lines of constant potential density because it requires no change in energy (adiabatic). Grey outlines and numbers denote different intermediate and deep water masses and their anthropogenic carbon content in PgC. From pmel.noaa.gov.

Calcifying organisms such as corals, coccolithophores, foraminifera, and pteropods build shells and other body parts out of calcium carbonate ($CaCO_3$) by using calcium Ca^{2+} and carbonate CO_3^{2-} ions (Fig. 14).

Figure 14: Examples of Calcifying Organisms.
Top left:(inactive link as of 5/19/2021) Coccolithophores (phytoplankton).
Top right: A live foraminifera (zooplankton).
Bottom left: Corals (animals).
Bottom right: (inactive link as of 5/19/2021) A pteropod, a.k.a. sea butterfly.



Decreasing carbonate ion concentrations and pH will lower the saturation state of calcium carbonate, which will make it more difficult for organisms to build calcium carbonate shells. It will also more easily dissolve existing calcium carbonate. Many scientists are concerned that the currently ongoing changes in the carbonate chemistry of the ocean and those expected for the future, in case of continued anthropogenic carbon emissions, may have adverse consequences for the ocean's ecosystem. The rates of change are likely much larger than anything experienced in the last millions of years, with unknown risks.

Experiments show that increased CO_2 or decreased pH can lead to malformed or partially dissolved coccoliths or pteropod shells. However, ocean acidification research is still in its infancy and consequences for many species and ecosystems are currently not known.

Questions

- What are Earth's three rapidly exchanging carbon reservoirs?
- Compare anthropogenic carbon emissions from the burning of fossil fuels with the natural carbon emissions from all volcanoes on Earth. How much different is one compared with the other?
- How much carbon does the average American emit per year?
    - We can visualize this amount by calculating the corresponding volume of CO_2. For this calculation we need to know that density $\rho = M/V$ is mass per volume. We also need to know the density of CO_2, which is $\rho_{CO2} = 2\ kg/m^3$, and we need to consider that one gram of carbon equals 3.7 grams of CO_2: 1 gC = 3.7 gCO_2. (That is because carbon has an atomic mass of 12, oxygen has an atomic mass of 18 and thus CO_2 has an atomic mass of 12+2×16 = 44. This makes carbon dioxide 44/12 = 3.7 times heavier than carbon.)
    - With the above information you should be able to calculate the volume V of CO_2 emitted by the average American.
    - If this volume $V = L^3$ is a cube, what is the length L of one of its sides?
    - How much carbon does the average American emit in a day?
    - You can visualize this amount as a volume of CO_2 or in bags of charcoal. A large bag sold in the US is about 20 pounds. You can assume charcoal is 100% carbon.

Videos

Lecture: Carbon Cycle 1
Lecture: Carbon Cycle 2
Lecture: Carbon Cycle 3

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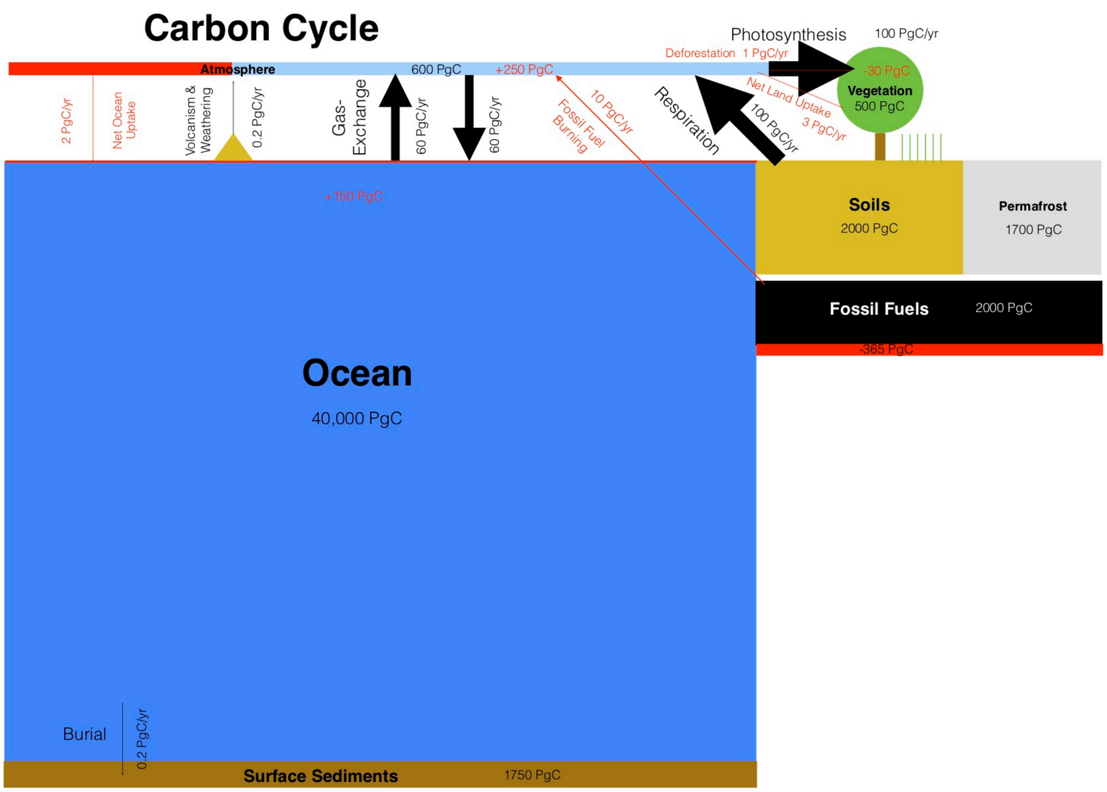
Additional Material



This figure of the global carbon cycle uses the same numbers as Fig. 1 but shows box sizes proportional to the size of the respective carbon reservoir. Red colors indicate human perturbations. Arrows are also proportional to the size of the fluxes, except for the small fluxes of sediment burial, volcanism and weathering, net ocean and land uptake and deforestation because those arrows would be hardly visible if drawn to scale. Based on Fig. 6.1 in IPCC (2013). [Here]() is a cool animation of changes over time.

6. Processes

Many different processes play important roles in the climate system. In chapter 4 we have already learned about radiation and the global energy budget. Here we want to discuss atmospheric and oceanic circulations, how they transport heat and water, and we'll explore a little more in depth Earth's water cycle and how it penetrates all climate system components and is linked to the energy cycle.

a) Atmospheric Circulation

Annual mean surface temperatures on Earth range from less than −40°C in Antarctica to +30°C in the tropics (Fig. 1). This begs the question: why is it warmer in the tropics than at the poles? The answer is, of course, because there is more sunlight in the tropics.

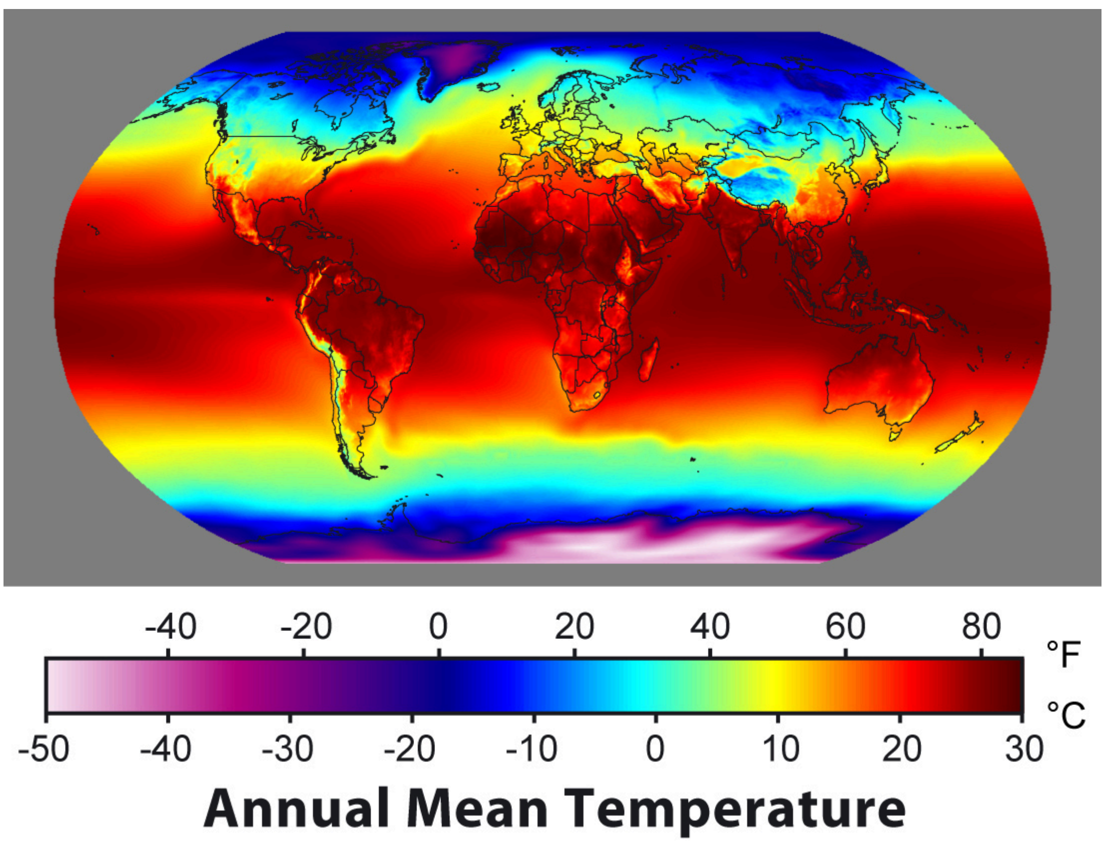


Figure 1: Annual mean surface temperature distribution. From wikimedia.org.

Due to the curvature of Earth's surface the equator receives more incident solar radiation per area than the poles (Fig. 2). Most solar radiation is received on a surface perpendicular to the sun's rays, whereas the more tilted Earth's surface is with respect to the sun's rays the less energy it receives. This is similar to holding a flashlight perpendicular to a surface or at an angle. The area lit is smaller if the flashlight shines perpendicular onto the surface. This configuration maximizes the energy input. The area lit is larger the larger the angle between the flashlight and the surface. In the extreme case of a 90° angle, or more, there is no light at all received. This situation corresponds to the poles or the dark side of the Earth.

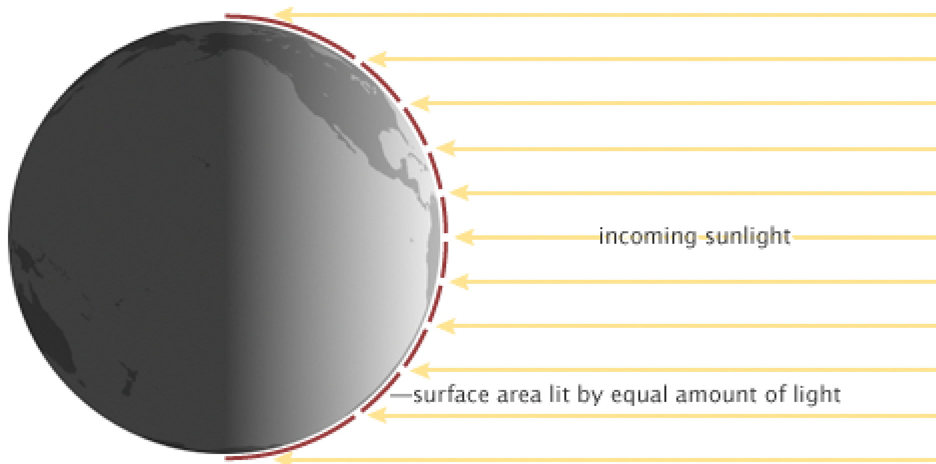


Figure 2: The amount of incoming sunlight on Earth is spread over a larger area at the poles than at the equator. This leads to less heating of the poles compared to the equator. From earthobservatory.nasa.gov.

Satellites have measured radiative fluxes at the top-of-the-atmosphere. Those data can be used to calculate the heat gain from absorbed solar radiation (ASR) and the heat loss from emitted terrestrial radiation (ETR) as a function of latitude (Fig. 3). The data show ASR values of around 300 Wm^{-2} in the tropics and 60 Wm^{-2} at the poles. Thus, the equator-to-pole difference is about 240 Wm^{-2}. However, values for the ETR are only around 250 Wm^{-2} in the tropics and between 150 and 200 Wm^{-2} at the poles. Thus, the equator-to-pole difference is only about 50 to 100 Wm^{-2}. The difference between the absorbed and emitted radiation gives the net heat gain from these fluxes. In the tropics the gain is positive. That is, there is more energy gain than energy loss. At the poles, on the other hand, there is more energy loss than energy gain. Therefore, if no other processes were involved the equator would warm up and the poles would cool down. But this is not the case, which implies a heat transport from the tropics towards the poles.

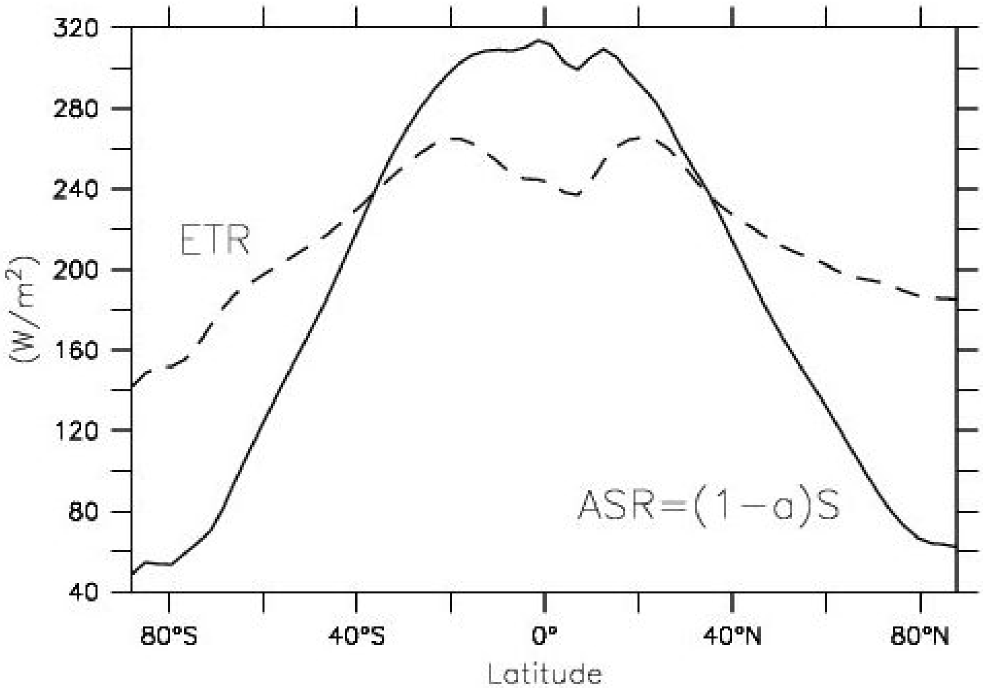


Figure 3: Absorbed solar radiation (ASR, solid) and emitted terrestrial radiation (ETR, dashed) zonally averaged as a function of latitude from satellite data of Earth's Radiation Budget Experiment.

Taking the difference ASR – ETR and integrating it from one pole to the other gives the total meridional heat flux (Fig. 4). In the southern hemisphere values are negative indicating southward heat transport. Positive values in the northern hemisphere represent northward transport. Poleward heat fluxes are peak at mid-latitudes with values of between 5 and 6 PW.

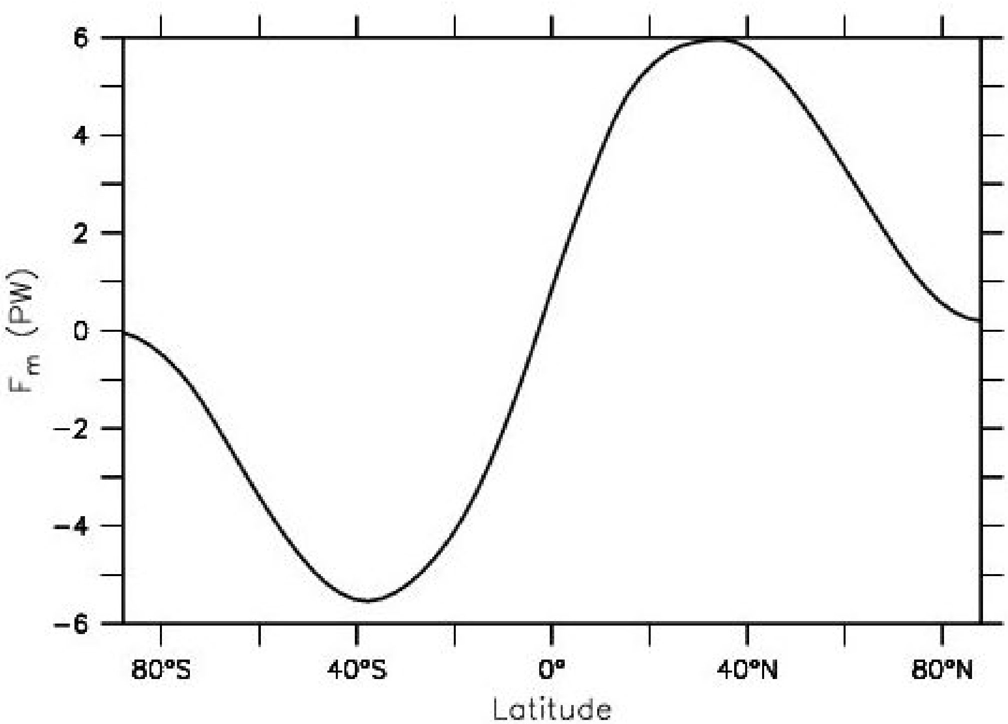


Figure 4: Meridional (northward is positive) heat transport in the climate system as a function of latitude implied from integration of the differences in the fluxes shown in Fig. 3. Units are peta watts (1 PW = 10^{15} W).

Most of the meridional heat transport is carried by the atmosphere (4-5 PW) whereas the ocean is responsible for a smaller portion (1-2 PW). Complex and turbulent motions in the atmosphere such as seen in [visualizations from satellites]() play an important part in this heat transfer.

Air is compressible. The weight of overlaying air compresses the air beneath and increases the pressure closer to the surface (Fig. 5). Therefore, pressure decreases with height and the surface pressure depends on the density and mass of the air above.

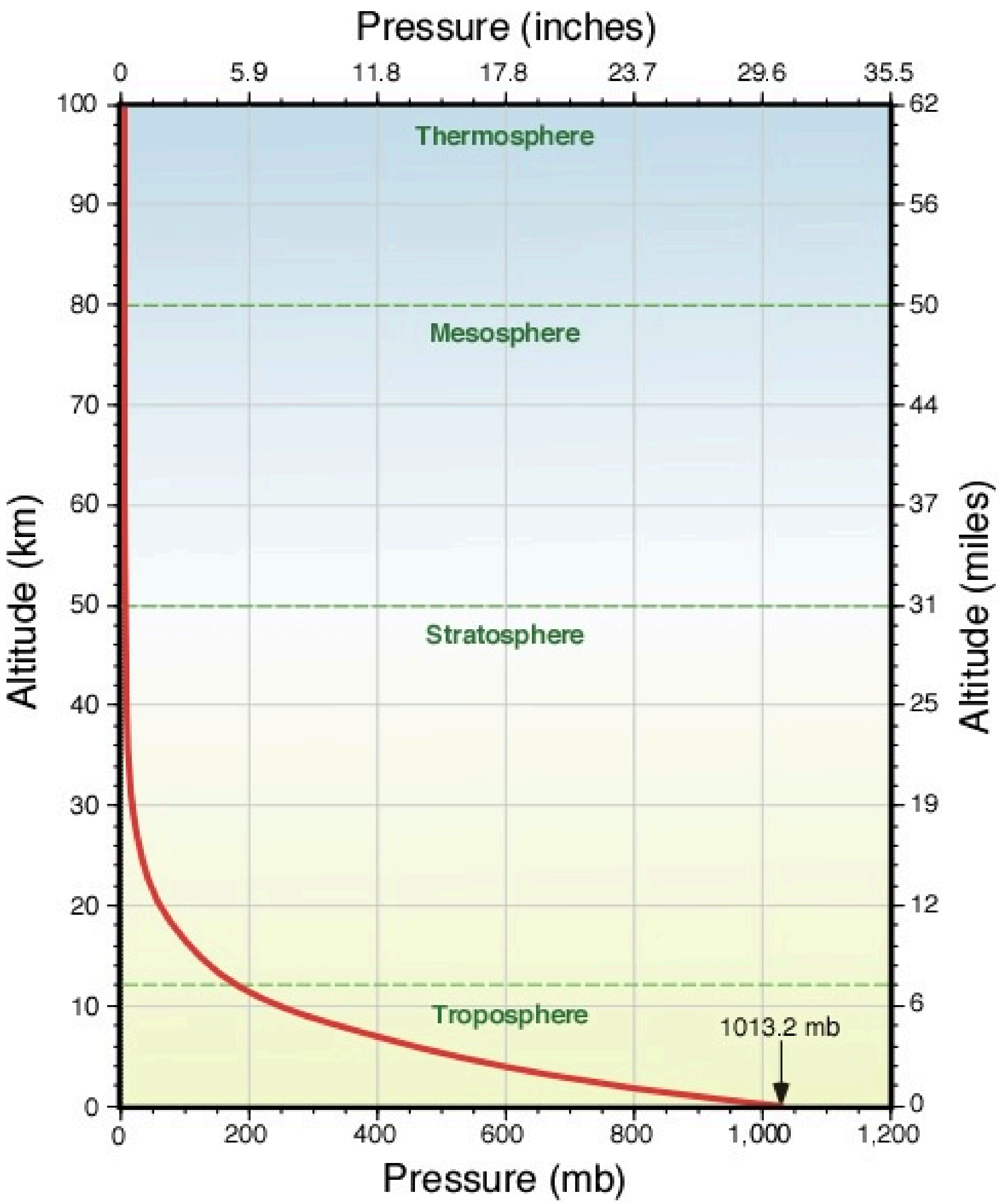


FIgure 5: Air Pressure as a Function of Altitude. From physicalgeography.net.

Let's imagine a non-rotating Earth (Fig. 6). In this case, the air at the poles would be cold and the air at the equator would be hot. Since colder air is denser than warm air the pressure at the north pole would be larger than at the equator. The air at the equator would rise and the air at the poles would sink. At the surface, the air would flow from high to low pressure, thus from the poles to the equator. At high altitudes the air would move from the equator to the poles.

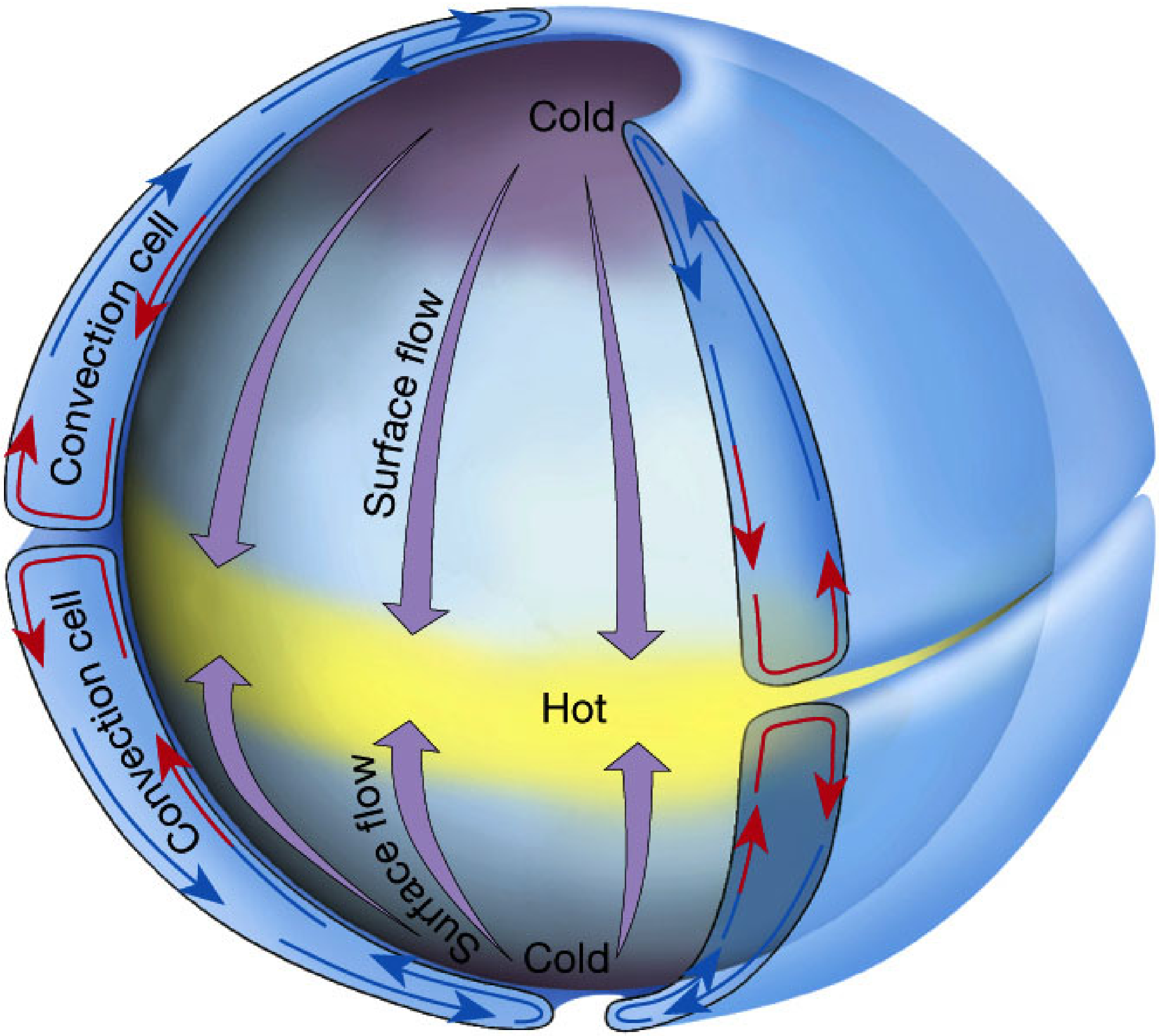


FIgure 6: Atmospheric Circulation on a Non-Rotating Earth. Figure 7.5 in The Atmosphere, 8th edition, Lutgens and Tarbuck, 8th edition, 2001. From [ux1.eiu.edu](ux1.eiu.edu).

However, Earth does rotate, which creates the Coriolis effect. Earth's rotation causes a deflection of air and water masses towards the right (left) in the northern (southern) hemisphere. The Coriolis force is a consequence of angular momentum conservation. Assume an air parcel at rest at the equator. (An analogy would be a spinning ice dancer with extended arms.) Its angular momentum is M = R×U, where U = 40,000 km/day = 500 m/s is approximately the velocity of the Earth at the equator. Now move the air northward. (The ice dancer pulls his/her arms in.) This will decrease R. In order to conserve angular momentum U has to increase. If the air moved to about 60°N its distance from the axis of rotation would have decreased by about one half. Therefore, U must have doubled. Thus, the air parcel would have a velocity of 500 m/s relative to the Earth's surface. Such high speeds never occur on Earth because of friction and turbulence but this simple example still explains qualitatively the high eastward wind velocities of around 40 m/s observed in the mid-latitude jet streams (Fig. 7).

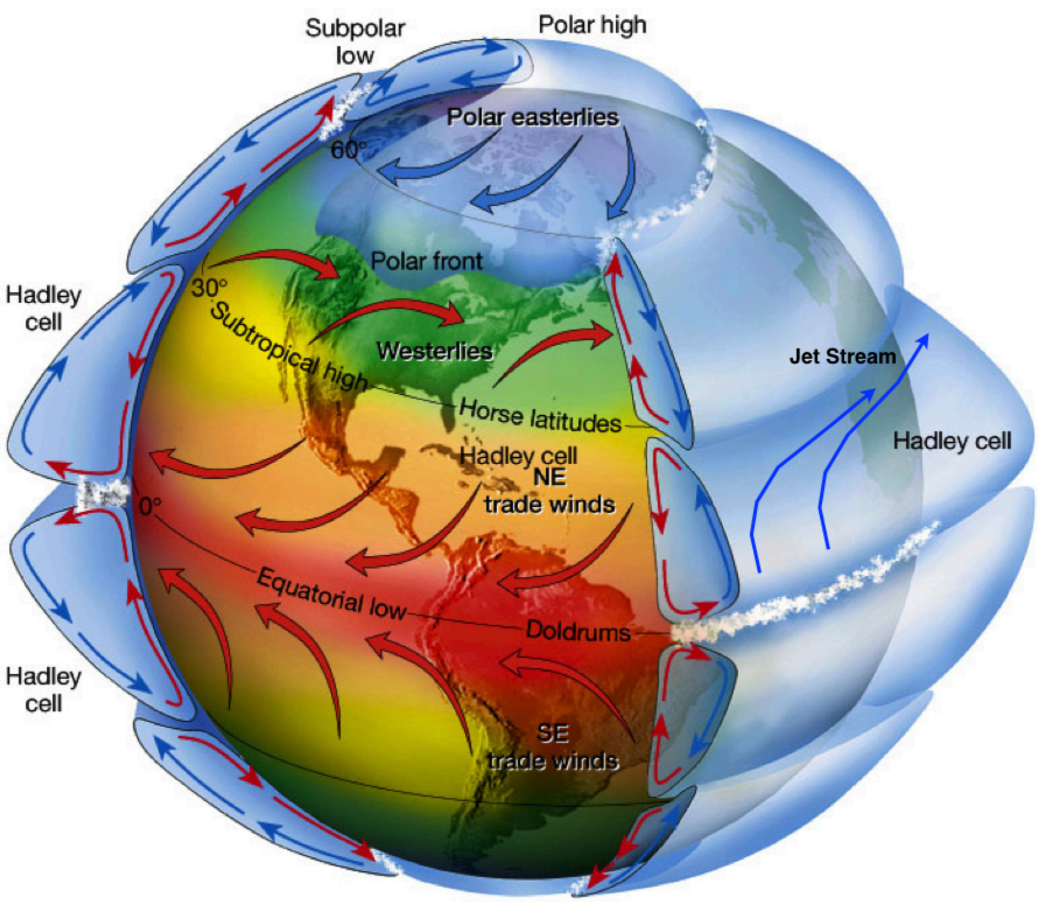


Figure 7: Atmospheric Circulation on a Rotating Earth. Modified from [ux1.eiu.edu](ux1.eiu.edu).

Rising of warm moist air at the equator causes water vapor condensation due to cooling of the air during the ascent. Clouds form and precipitation occurs. Some of the deepest cumulonimbus clouds on Earth form in the tropics. They can reach the top of the troposphere or higher. The cool relatively dry air then moves poleward. Now the Coriolis effect kicks in, deflects the air towards the right (left) in the northern (southern) hemisphere, which creates the jet stream. The air cools by emitting longwave radiation to space. This increases the density and the air descends back to the surface in the subtropics (~30°N/S). During the descend the air warms and its relative humidity decreases. This leads to dry conditions in the subtropics indicated by the major deserts at those latitudes.

Subsequently the dry air moves back towards the equator. The Coriolis force deflects it towards the right (left) in the northern (southern) hemisphere, creating the easterly trade winds in the tropics. During this movement along the sea surface the air picks up water vapor from evaporation. Once the air returns to the equator it is saturated with water vapor (close to 100% relative humidity). The resulting meridional overturning cells in the tropical atmosphere are called Hadley cells, or Hadley circulation. Two cells, one in each hemisphere, exist only during the fall and spring, whereas during summer/winter there is only one major cell with rising air just slightly off the equator in the summer hemisphere, where the heating is largest.

The belt of rising air close to the equator is called the Intertropical Convergence Zone (ITCZ), due to the convergence of air along the surface. The ITCZ is further north in the northern hemisphere summer and further south in the southern hemisphere summer, although on average it is slightly north of the equator because the northern hemisphere is slightly warmer than the southern hemisphere due to ocean heat transport from the southern to the northern hemisphere (Frierson et al., 2013).

Water in the Hadley Cell

Let's follow an air parcel of about 1 kg weight (~1 m^3 at the surface) during its travels along the Hadley cell and estimate its water vapor content using Fig. 16 of chapter 4.

- Starting at the ITCZ we assume the temperature is 30°C and the air is fully saturated with water vapor. How many grams of water vapor does it contain?
- The air ascends to the top of the troposphere. It cools to about -30°C. It is still at saturation. How many grams of water vapor does it contain?
- How many grams of water vapor has the air parcel lost?
- Now the air parcel moves poleward and descends in the subtropics. The descend causes warming. Will the water vapor content change during the descend?
- Let's assume the surface temperature is close to 30°C. What will be the relative humidity of the air?
- During its travels near the surface evaporation from the ocean will quickly increase the air parcel's water vapor content close to saturation. How many grams of water vapor will the air parcel have gained?

Surface air moving from the high pressure at subtropical latitudes towards lower pressure at mid-latitudes also experiences the Coriolis effect. This leads to the prevailing westerly winds at mid-latitudes. Another important feature of the atmospheric circulation at mid-latitudes is the growth, movement, and decay of synoptic weather systems (transient eddies) that dominate weather variability and heat transport there. Transient eddies are the low and high pressure systems that move eastward, some of which can be associated with storms.

Explore these features of the global atmospheric circulation in this animation of weather variations throughout a whole year from satellite observations. You may notice the pulsing of convection over tropical Africa and South America. This is caused by the diurnal (daily) cycle of surface heating. Fig. (8) shows the observed global distribution of precipitation. Note the ITCZ as the band of high precipitation close to the equator, the regions of low precipitation in the subtropics and the bands of high precipitation at mid-latitudes in the paths of the storm tracks over the North Pacific and North Atlantic. In the southern hemisphere we see an additional band between 50-60°S.

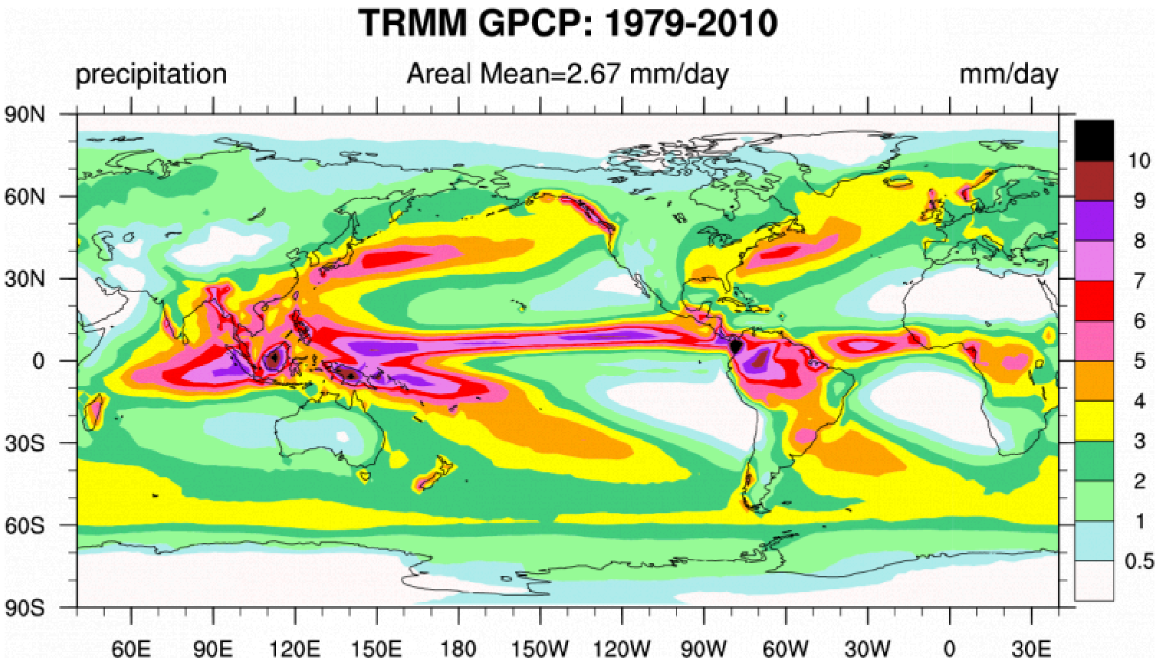


Figure 8: Annual mean precipitation estimated from observations. From ucar.edu.

Precipitation has a strong effect on vegetation as can be seen from the similarities between Figures 8 and 9. Regions of large precipitation such as tropical Africa, South America, and Asia have dense vegetation, whereas regions of little precipitation such as the Sahara, the Arabian Peninsula, regions in central Asia, southwestern parts of North America, central and western Australia, parts of South America west of the Andes, and southwestern Africa also have sparse desert or steppe vegetation. This relation is not a surprise considering that photosynthesis requires water (see box Photosynthesis and Respiration in chapter 5).

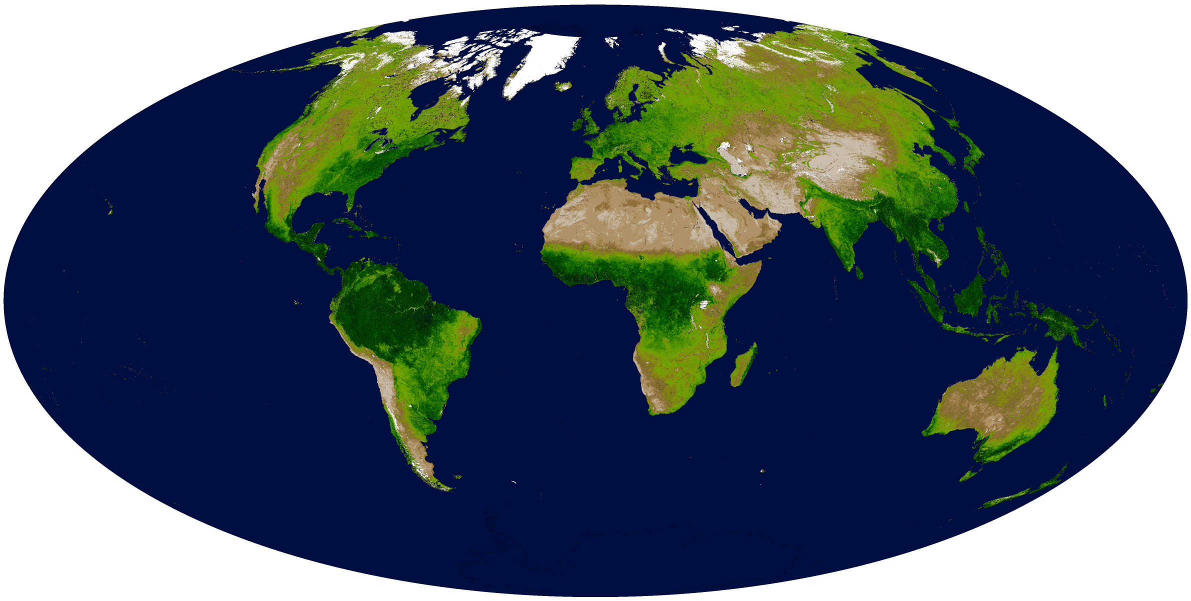


FIgure 9: Global Vegetation Distribution From Satellites. Shown is the Enhanced Vegetation Index (EVI), which represents the concentration of green leaf vegetation. From nasa.gov.

b) The Hydrologic Cycle

Water plays a fundamental role in the climate system. It is involved in Earth's energy cycle and links physical and biological processes. Water has some remarkable properties due to its molecular structure. Hydrogen bonds in liquid water result from attractive electric forces between the positively charged hydrogen atoms of one molecule with the negatively charged oxygen atom of a neighboring molecule. A large amount of energy input is required to overcome this force for a transition from the liquid to the vapor phase (Fig. 10). This energy is called **latent heat** of vaporization. It is about 2,300 joules per gram of water. The same amount of energy is released during condensation.

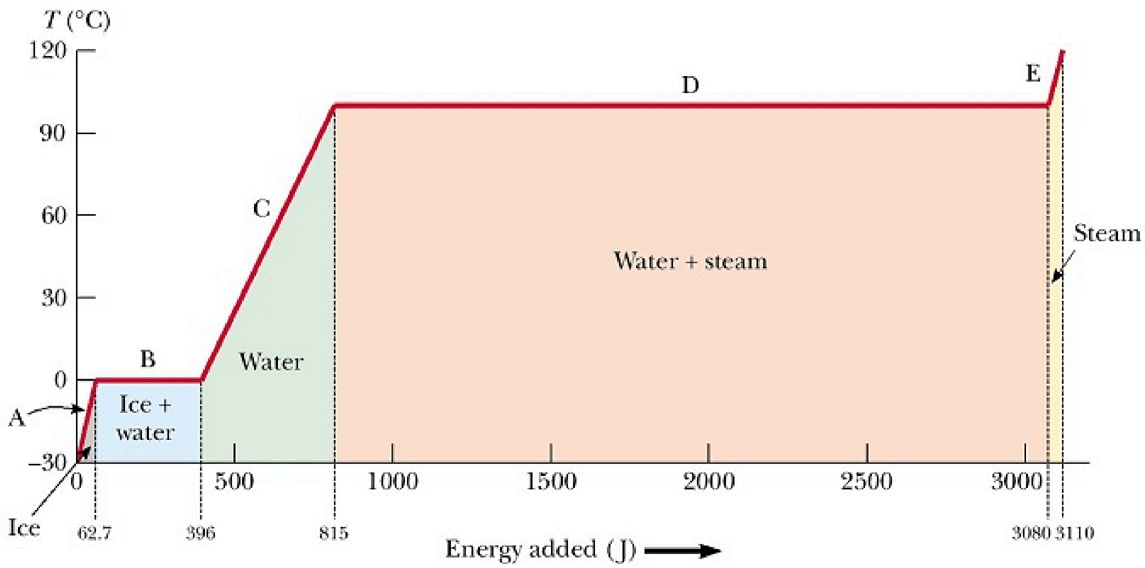


Figure 10: Effects of heat added (ΔH) on the temperature of one gram of water. Starting from ice at -30°C we need to add about 63 J to warm it up to the melting point (line A). About 333 J are required to melt the ice (line B). During this process the temperature stays constant. Once all ice is melted into water adding more heat causes the water temperature to increase until the boiling point (line C). The slope of line C is the heat capacity of water (at constant pressure) $c_{p,water}$ = ΔH/ΔT = 4.2 J/(g°C). Once the water boils any added heat is used to vaporize water and the temperature stays at the boiling point until all water is vaporized. The latent heat of vaporization is ~2270 J/g. Thus, the energy required to vaporize water is more than five times as much as to warm it from 0°C to 100°C.

Lapse Rate

The lapse rate Γ = ΔT/Δz is the rate of temperature decrease ΔT with height Δz in the atmosphere. Let's try to estimate the effect of condensation of water vapor in the ascending branch of the Hadley cell on the temperature of the upper atmosphere (z = 10 km). In the previous box we have calculated that approximately 30 g of water vapor were lost from 1 kg of air during its ascent.

- How much latent heat of condensation was released?
- How much would that added heat have increased the temperature of the air parcel, assuming a specific heat capacity of air of $c_{p,air}$ = 1 J/(g°C)?

The observed lapse rate of the atmosphere is on average about Γ_m = -6.5 °C/km, which is close to the moist adiabatic lapse rate. (Adiabatic means that no heat is added or removed from the air parcel.) This is in contrast to the dry adiabatic lapse rate, which is approximately Γ_d = -10 °C/km. Thus, given a surface air temperature of 30°C, the air at 10 km altitude at the equator would be -70°C in a dry atmosphere compared with -35°C in the real, moist atmosphere. Our calculation above was not quite correct due to the various assumptions we made, but the order of magnitude was right. This example illustrates the large effect of latent heat release on upper air temperatures.

Condensation occurs when the air is at 100% relative humidity and when cloud condensation nuclei are available. Cloud condensation nuclei are small particles in the air. In the atmosphere condensation typically occurs when the air is cooled, e.g. during ascending motions in convection or when air is lifted over mountains. The latent heat released during condensation in clouds leads to warming and thus more intense upward motions. This is an important driver of convection and storms (Fig. 11).

Figure 11: Picture of a convective cloud. Warming from latent heat release during condensation makes the interior of a cloud more buoyant than the surrounding air and thus intensifies convection. From globalwaterforum.org.

Cloud in a Bottle

You can make your own cloud in a bottle and experience the effects of condensation nuclei for yourself. All you need is a plastic bottle, water and matches.

1. Take a large plastic bottle and fill it to about 3/4 with water.
2. Screw on the lid tight.
3. Squeeze the bottle real hard for a few seconds and then stop squeezing.

What do you see? Nothing? Don't worry that's normal. Your cloud cannot form because it lacks condensation nuclei. Now, open the lid, light a match, blow it out and throw it in the bottle. The smoke particles will act as condensation nuclei. Repeat steps 2 and 3 of the experiment and a cloud should occur. By squeezing the bottle you increase the pressure and thus the temperature of the air, which causes more evaporation. As you release the pressure, the temperature drops and the water vapor condenses on the smoke particles. Here is a video of the experiment.

Evaporation occurs when the relative humidity $rh = q/q_{sat}$ of the air above a water surface is less than 100%. q is the specific humidity, that is the amount of water vapor (in grams) per moist air (in kg) and q_{sat} is the specific humidity at saturation. The lower the relative humidity the higher the rate of evaporation $E \sim q_{sat} - q$. Stronger winds also cause more evaporation, similar to your blowing over your hot coffee or tea to cool it down. Evaporation leads to cooling of the remaining liquid water since it removes the fastest molecules, and the slower ones stay behind. This principle is also at work in air conditioners and refrigerators, in which air is cooled by evaporation of a coolant. Evaporative cooling is important in keeping Earth's surface and especially the ocean cool. Over vegetated land areas **transpiration** of water by plants also cools the surface. Plants can limit their water loss through transpiration by closing their stomata.

Another property of water that is important for climate is its large **heat capacity**. The table below compares the heat capacity of water with that of air. On a per gram basis water already has more than four times the heat capacity of air. Moreover, the density of water is 1000 times that of air. Therefore, on a per volume basis the heat capacity of water is 4200 times that of air. As a result, the top 2 m of the ocean has the same heat capacity as the whole atmosphere.

		Units	Air	Water
Specific Heat Capacity	c_p	J/(g°C)	1	4.2
Density	ρ	kg/m^3	1	1000
Volumetric Heat Capacity	$c_{vol} = c_p\rho$	$J/(m^3°C)$	1000	4,200,000

Experiment: Heat Capacity

The differences in heat capacity of air versus water can be demonstrated in a simple experiment with two balloons. You can do this at home. Fill one balloon with with air, the other with water. We will hold the flame of a lighter to the balloon.

- But first make a guess what will happen. Will there be a difference in the results?
- Now perform the experiment with the air balloon first. What happened?
- Now perform the experiment with the water balloon. Is there a difference?
- Did you expect these results?

If you don't want to do the experiment yourself you can watch a video here.

Land also has a much lower heat capacity than the ocean. Let's consider the budget equation for two cases: an air column with ocean and one with land. Let's also assume the system was initially in balance, e.g. $\Delta CT/\Delta t = I - O = 0$ (zero), which means the rate of change in heat content CT is zero and thus the temperature T is constant. C is the heat capacity, which is larger for the column with the ocean underneath. Now, we add a forcing ΔF on the right-hand-side such that $\Delta CT/\Delta t = \Delta F$. Because the heat capacity is a constant (it does not change with time) we can divide by C to get the temperature change $\Delta T/\Delta t = \Delta F/C$. This equation implies that the temperature change over the ocean will be slower than over land because it has a larger heat capacity C.

The effects of the differences in heat capacity between the ocean and land can be seen in Fig. 12, which shows the amplitude of the seasonal cycle in surface temperatures (summer minus winter). The seasonal cycle in the forcing is much larger at higher latitudes than in the tropics, which explains why the seasonal temperature variations in the tropics are smaller than at higher latitudes. But you also see a larger difference between the seasonal cycle over land and over the ocean at similar latitudes. E.g. over the North Pacific at 40°N temperatures in summer are only about 10°C warmer than in winter, whereas in the interior of North America they are 30°C warmer. The largest amplitudes of the seasonal cycle are over east Siberia, because it is the furthest downstream from the ocean (remember the prevailing westerly winds at those latitudes) and over Antarctica because it is, for dynamic and topographic reasons, isolated. Dynamically the large westerly winds over the Southern Ocean inhibit meridional transport. Topographically, the sheer height of the ice sheet removes it from what is happening at the sea surface.

In the data explorations in chapter 1 you may have found similar features such as a smaller seasonal cycle closer to the ocean than in the continental interior. Most people live close to the ocean presumably at least partly because climate variations there are more dampened and less extreme than in the interior of the continents.

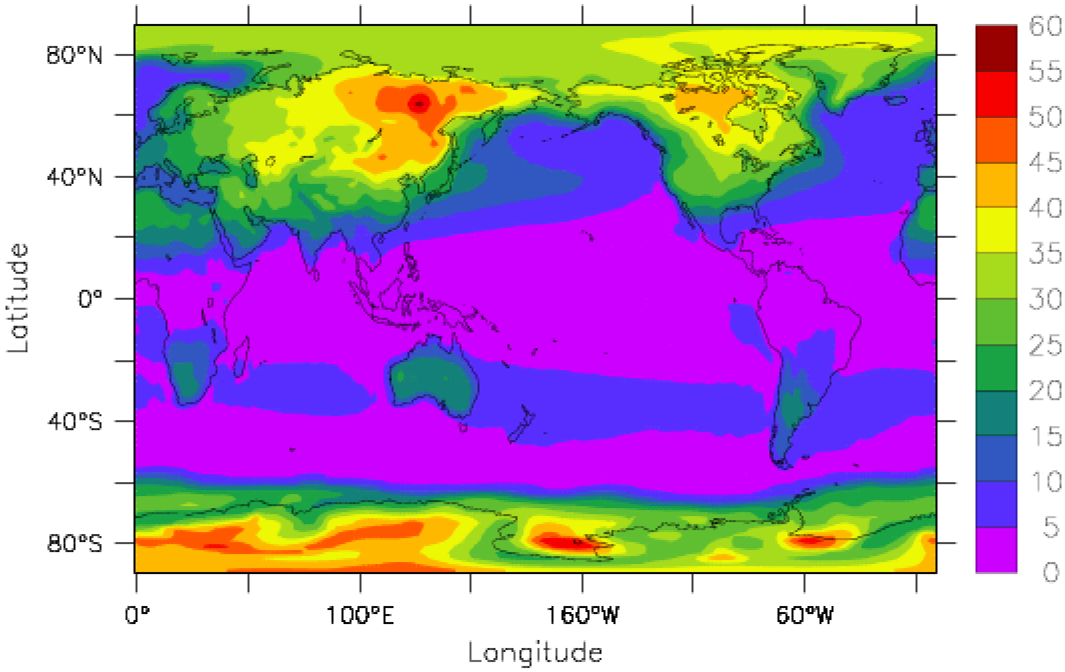


Figure 12: Amplitude of the Seasonal Cycle of Surface Temperatures. Plotted is the maximum minus the minimum of the average seasonal cycle from 1958 to 1996 from the National Center for Environmental Prediction (NCEP) Reanalysis. Units are in degrees Celsius.

The large heat capacity of the ocean not only dampens seasonal temperature variations but also those on other timescales. We have seen in chapter 2 (Fig. 2) that observed warming over the past 100 years is also smaller over the ocean than over land. Now we understand that the observed land-sea contrast is due, at least in part, to the differences in heat capacities.

In fact, about 90% of the observed increase in Earth's heat content goes into the ocean (Fig. 13). Most of the remaining 10% goes into ice and land, whereas the heat gain of the atmosphere is very small in comparison.

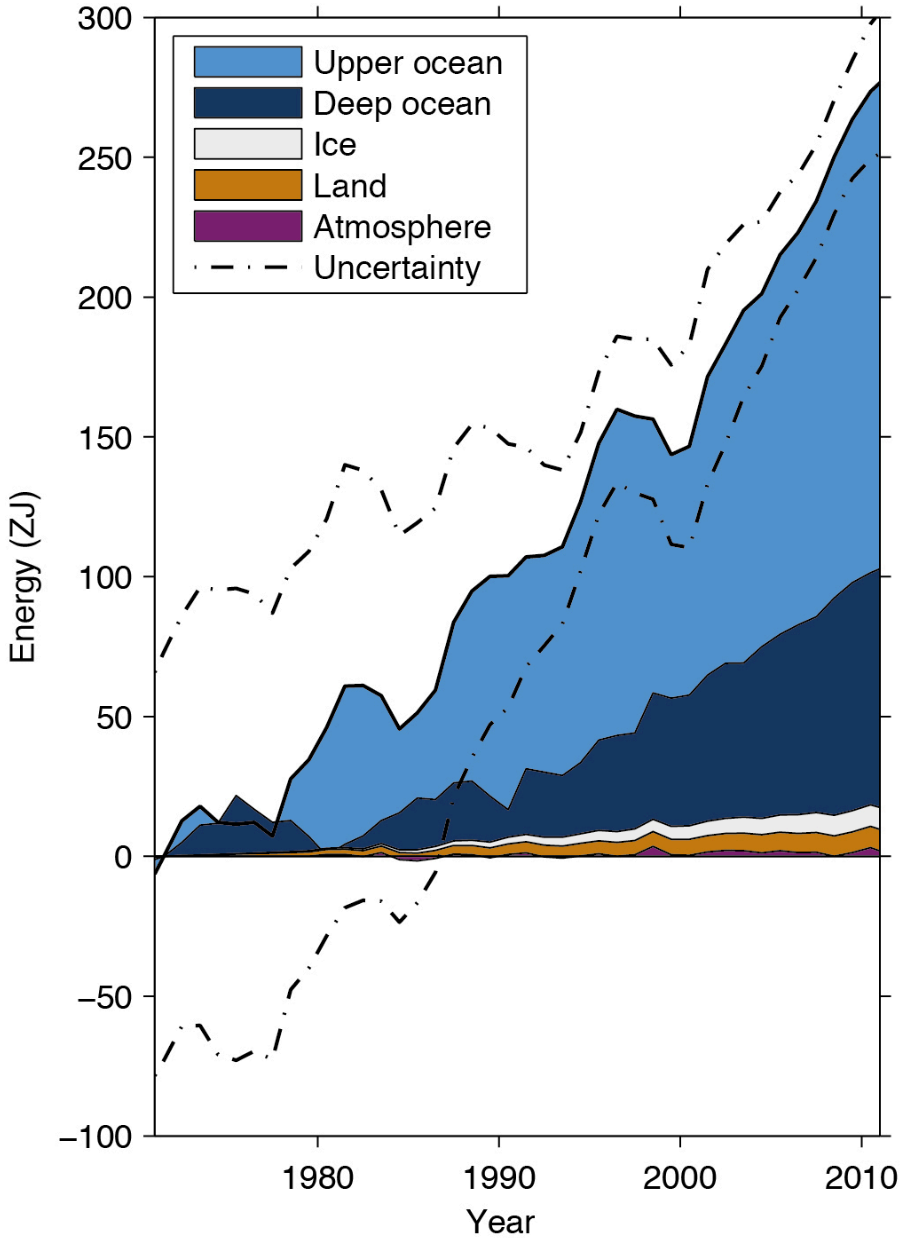


FIgure 13: Changes in Earth's Heat Content. From ipcc.ch.

Fig. (14) shows a schematic of the global water cycle. Most water is contained in the ocean (more than one billion cubic km), whereas the atmosphere contains only a relatively small amount (13 thousand cubic km). Evaporation removes about 400 thousand cubic km from the ocean each year, most of which precipitates back over the ocean. 40 thousand cubic km are transported in the atmosphere from the ocean to land each year. Over land this water precipitates together with about 70 thousand cubic kilometers of recycled water from evapotranspiration from land surfaces.

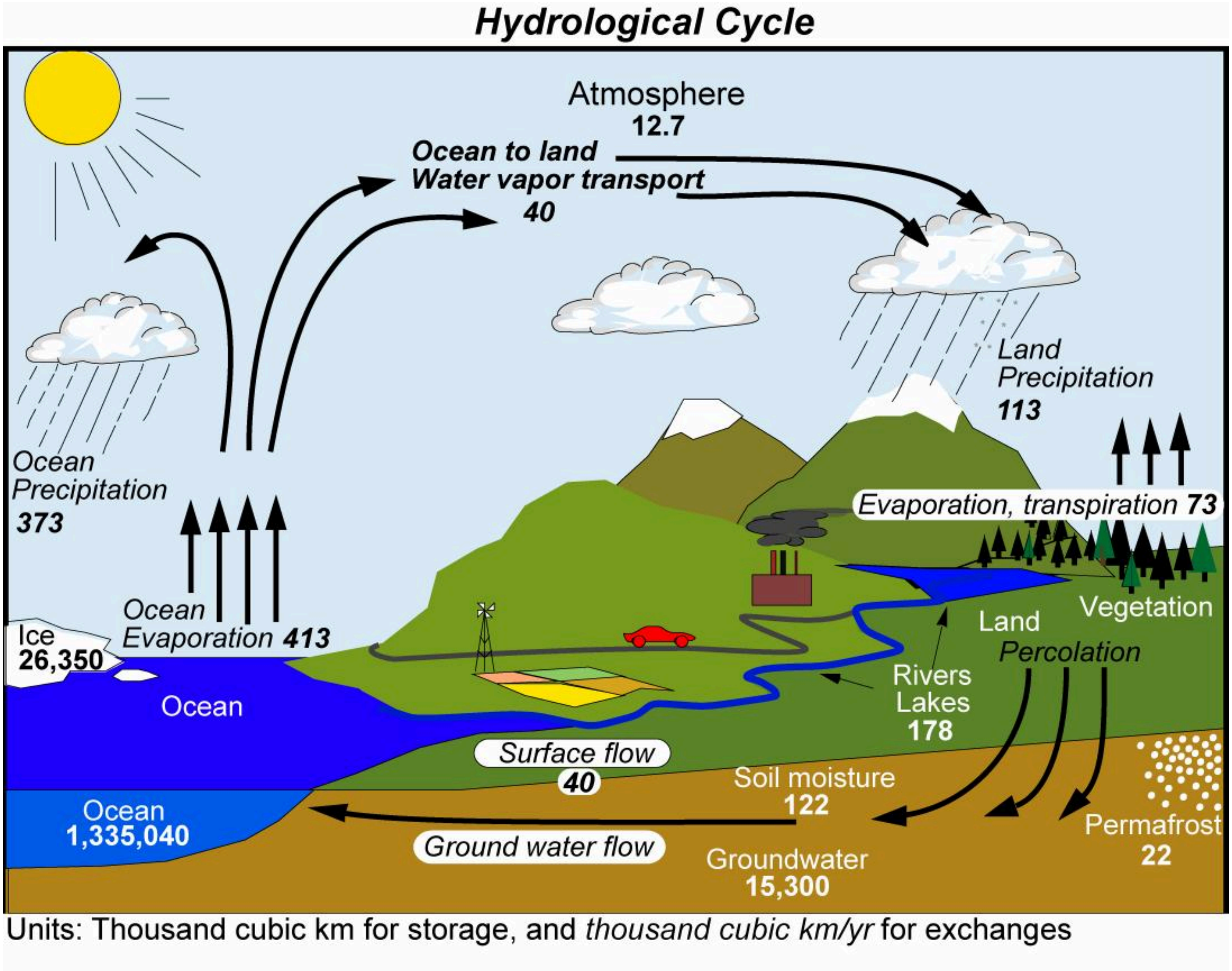


Figure 14: The Global Water Cycle. Fluxes are indicated in slanted font. From Trenberth et al. (2007). Image from ucar.edu.

Observations show increases in water vapor content in the atmosphere (Fig. 15). This is consistent with our understanding of the physics of the hydrologic cycle and its dependency on temperature (Clausius-Clapeyron relation). Warmer air can hold more water vapor, and due to the presence of the oceans there is no lack of water supply to the global atmosphere.

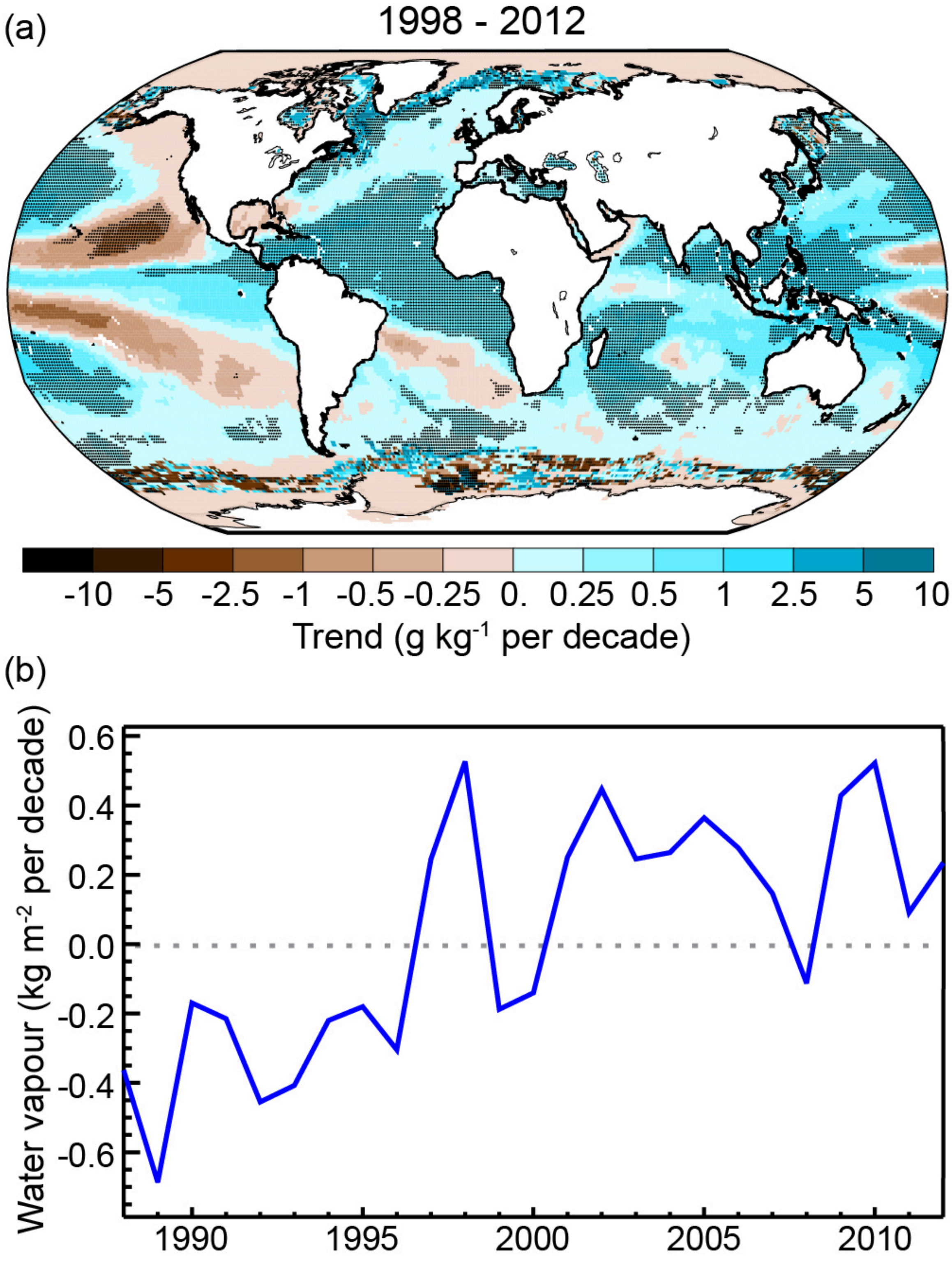


Figure 15. (a) Observed changes in surface specific humidity estimated from satellites. Dark shading indicates significant changes. (b) Changes in globally averaged water vapor content. From Harmann et al. (2013). Image from ipcc.ch.

c) Ocean Circulation

The general, planetary-scale circulation of the ocean can be separated into a wind-driven component that dominates the upper ocean and a density driven component that occupies the deep ocean. Five large gyres are the main features of the surface circulation in the subtropics (Fig. 16). The easterly trade winds push water towards the west in the tropics. The water piles up where it encounters land and flows poleward. The poleward flow brings warm waters from the tropics to the mid-latitudes. There, the westerly winds push the surface waters toward the east. Again, where the current hits a continent it piles up and flows north and south. The southward flowing part completes the subtropical gyre bringing cold waters towards the tropics. The poleward flow along the western boundaries of the subtropical gyres are warm currents, such as the Gulf Stream, the Kuroshio, and the Brazil Currents. Equatorward currents along the eastern boundaries are cold such as the California or the Peru (or Humboldt) Currents. Within the subtropical gyres the water flows in a spiral-like pattern towards the center of the gyre. This convergence causes sinking in the centers of the gyres. However, the water is relatively warm so it sinks only to depths of a few hundred meters. In the tropics, on the other hand, the trade winds cause divergence and upwelling.

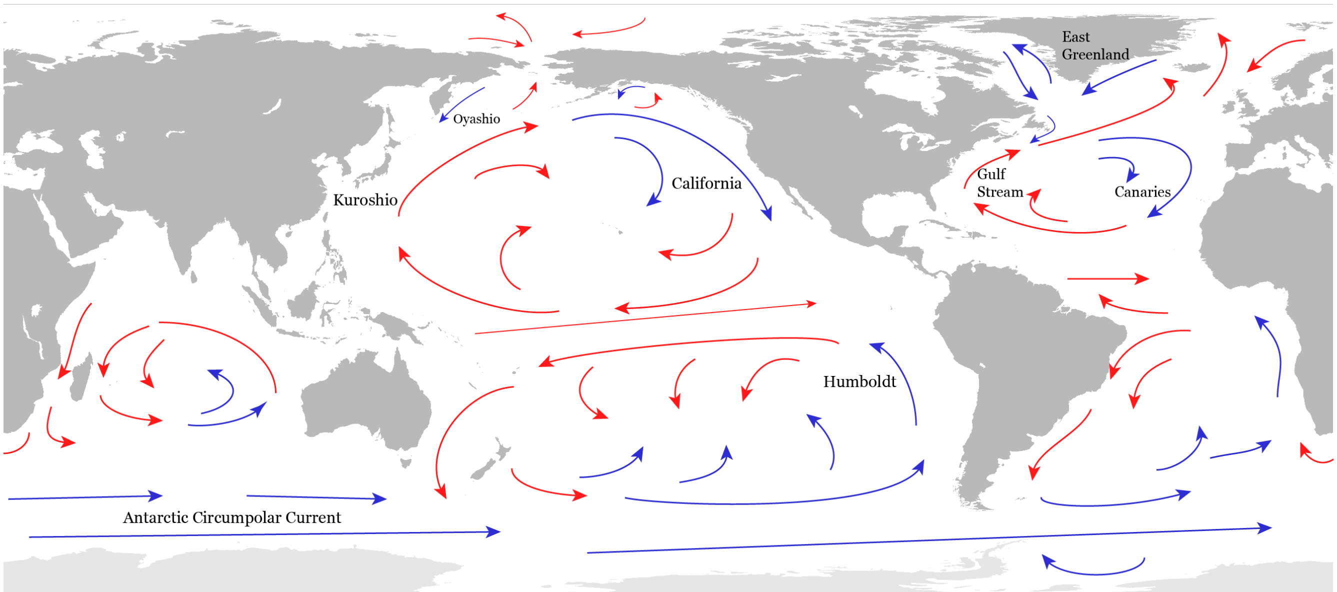


Figure 16: Sketch of the Surface Circulation of the Ocean. After Peixoto and Oort (1992). Warm or poleward flowing currents are depicted in red, cold or equatorward flowing currents in blue. Names for some currents are noted.

Surface currents near the equator are particularly vigorous. The piling up of waters in the western equatorial Pacific, for example, causes the narrow Equatorial Counter Current to move eastward.

The strongest current in the world oceans is the Antarctic Circumpolar Current, which transports more than 100 million cubic meters of water per second around Antarctica flowing eastward. That amounts to 500 times the Amazon river discharge. The Southern Ocean is also an important region for upwelling of deep waters caused by a wind-driven divergence of surface waters.

Watch a high resolution model simulation of surface ocean currents here. It shows more details such as mesoscale eddies, which are the ocean's equivalent of weather systems in the atmosphere. Notice that they are much smaller in size than high and low pressure systems in the atmosphere due to the larger density of seawater compared with air. Try to identify some of the features discussed above such as the Gulf Stream and the Antarctic Circumpolar Current. These and other fascinating features are also seen in satellite observations of sea surface temperatures here and here.

In contrast to the atmosphere, the ocean is mostly stably stratified. That is, denser water is layered below lighter water. The density of sea water is determined by temperature and salinity. The colder and saltier, the heavier it is. Typically warmer, more buoyant water is on top of colder water, especially at low latitudes (Fig. 17). This is because water absorbs sunlight efficiently, which heats the surface. Winds cause lots of turbulence close to the surface, which creates a layer of uniform temperature called the surface mixed layer. Below that, between about 200 – 1000 m depth, is a region in which the temperature decreases rapidly with depth. This is called the thermocline. Turbulence is weak here. Further down is the weakly stratified deep and abyssal ocean. Vertical temperature gradients are small here. Closer to the sea floor turbulence increases due to interactions of flow with the bottom topography.

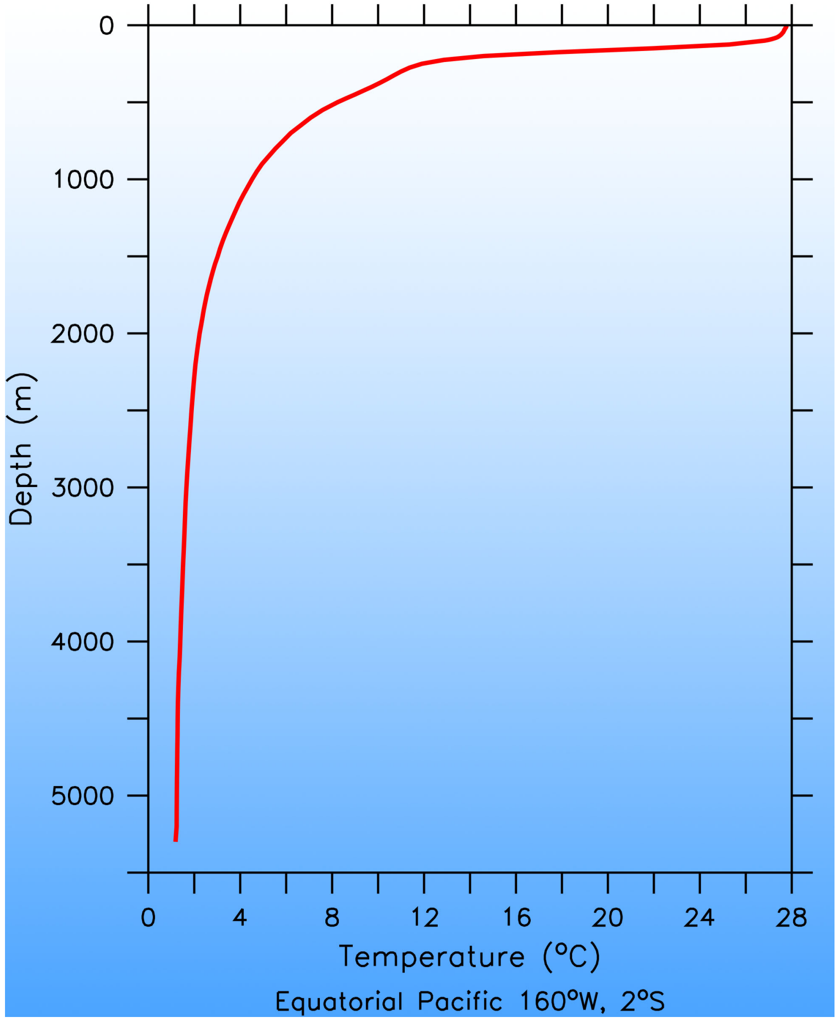


Figure 17: Typical Temperature Profile with Depth in the Ocean.

The deep ocean is cold because waters there originate from the high-latitude surface. Only in a few regions of the world's oceans where the density of surface waters is large enough do they sink into the deep ocean (Fig. 18). In the current ocean there is deep water formation in the North Atlantic and near Antarctica. Surface waters of the Atlantic are saltier than those of the Pacific because of water vapor transport within the atmosphere. Whereas mountain ranges at mid-latitudes block water vapor transport from the Pacific to the Atlantic with the westerly winds there, in the tropics gaps in the mountains allow water vapor transport with the trade winds from the Atlantic to the Pacific. This causes fresher, more buoyant surface waters in the Pacific. In the North Pacific this freshwater lens prevents sinking, whereas in the North Atlantic saltier waters are dense enough to sink to about 2-3 km depth. From there they flow south along the margin of the Americas pushed there by the Coriolis force. The deep water from the North Atlantic crosses the equator and the South Atlantic and enters into the Southern Ocean. Part of it rises back to the surface there whereas the rest flows into the Indian and Pacific Oceans, where it slowly ascends. The return flow at the surface flows through the Indonesian Archipelago into the Indian Ocean, merges with upwelled waters there and continues to flow westward around the tip of South Africa and back northward across the Atlantic. This planetary scale circulation pattern is called the thermohaline (thermo = temperature, haline = salinity) circulation or meridional (north-south) overturning circulation.

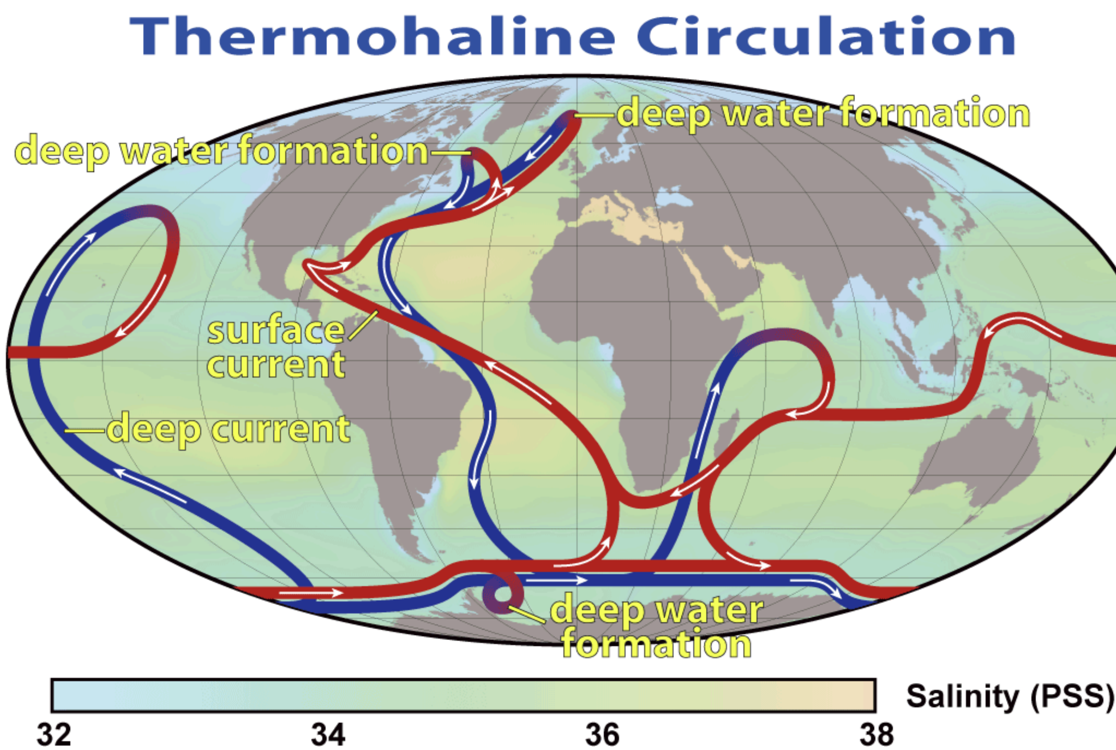


Figure 18: Cartoon of the Deep Ocean Circulation. Red and blue ribbons represent surface and deep currents, respectively. From wikipedia.org.

Its circulation impacts tracer distributions in the ocean (Fig. 19). E.g. in the Atlantic the southward flowing North Atlantic Deep Water (NADW) can be identified as a water mass with relatively high salinity between about 2 – 4 km depth. Fresher Antarctic Bottom Water (AABW) flows north below NADW. It is colder and therefore denser than NADW. Relatively fresh Antarctic Intermediate Water (AAIW) flows north above NADW creating a sandwich-like structure in the deep Atlantic. In the North Atlantic there is a blob of high salinity water around 1 km depth and 40°N. This is outflow from the Mediterranean Sea, which is very salty.

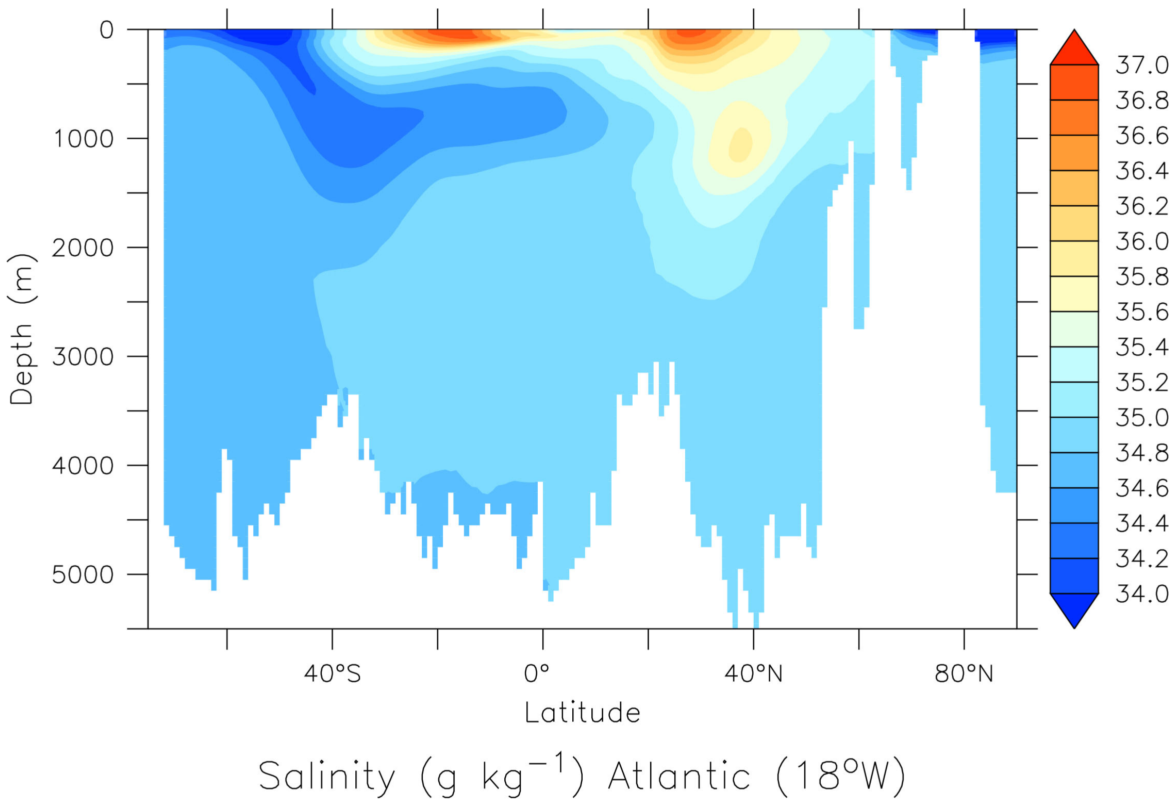


Figure 19: Salinity (in grams salt per kg of sea water) section through the Atlantic ocean at 18°W as a function of latitude and depth.

Experiment Thermohaline Circulation

Perform a simple experiment that illustrates the effects of salinity and temperature on the density of water. You'll need the following ingredients:

- a container, preferably made out of a transparent material such as glass or plexiglass,
- ice cubes,
- a small sponge, and
- food coloring.

Fill the container with water. Now put the moist sponge at the surface of the water so that it swims on the top. Pour some salt on the sponge. Not too much. You don't want it to spill over or capsize the sponge. Just enough so that the water in contact with the sponge will soak up the salt. Now add a few drops of food coloring on top of the salty sponge and observe what happens. Where does the water flow?

Now add an ice cube to the water. Drip a few drops of food coloring (choose a different color than before) onto the ice cube and observe. What happens? Describe your observations. Explain your observations with what you've learned about the effects of salinity and temperature on sea water density.

Observations show that the ocean is warming (Fig. 20). Most of the increase in temperatures is concentrated near the surface consistent with a warming atmosphere as its cause. A prominent maximum of heat uptake is in the North Atlantic, similar to the pattern of anthropogenic carbon uptake (Fig. 12 in chapter 5). The reason is the sinking and southward penetration of NADW, which transports both anthropogenic carbon and heat from the surface to the deep ocean there. Other regions of enhanced heat uptake are the subtropics, where surface waters sink to a few hundred meters depth in the centers of the subtropical gyres. Ocean temperature observations rule out that changes in ocean circulation are the cause of the observed warming of the surface. If this was the case deeper layers would have cooled, which is not what is observed. Therefore, the hypothesis that changes in ocean circulation have caused the observed warming of the atmosphere during the past 50 years has been falsified by subsurface temperature observations.

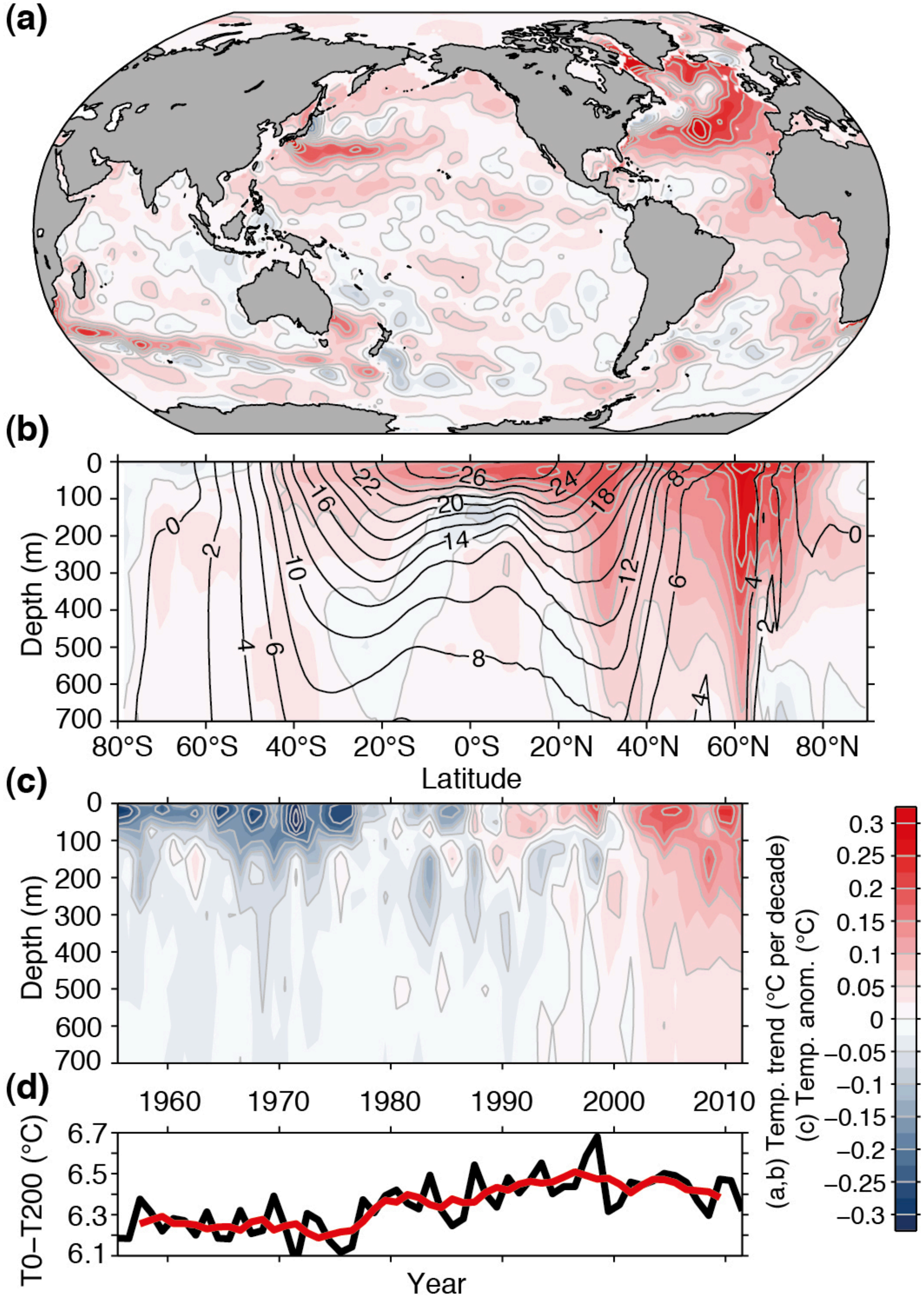


Figure 20: (a) Temperature trend over the top 700 m from 1971–2010. (b) Zonally averaged temperature trend. Black contour lines represent the mean temperature distribution. (c) Horizontally averaged temperature changes. (d) Changes in temperature difference between 200 m and the surface. From Rhein et al. (2013). Image from ipcc.ch.

Acceleration of the atmospheric hydrological cycle also affects ocean surface salinities (Fig. 21). Regions that are already salty such as the subtropics and the Atlantic get even saltier and regions that are already fresh like the North Pacific and the Southern Ocean get even fresher.

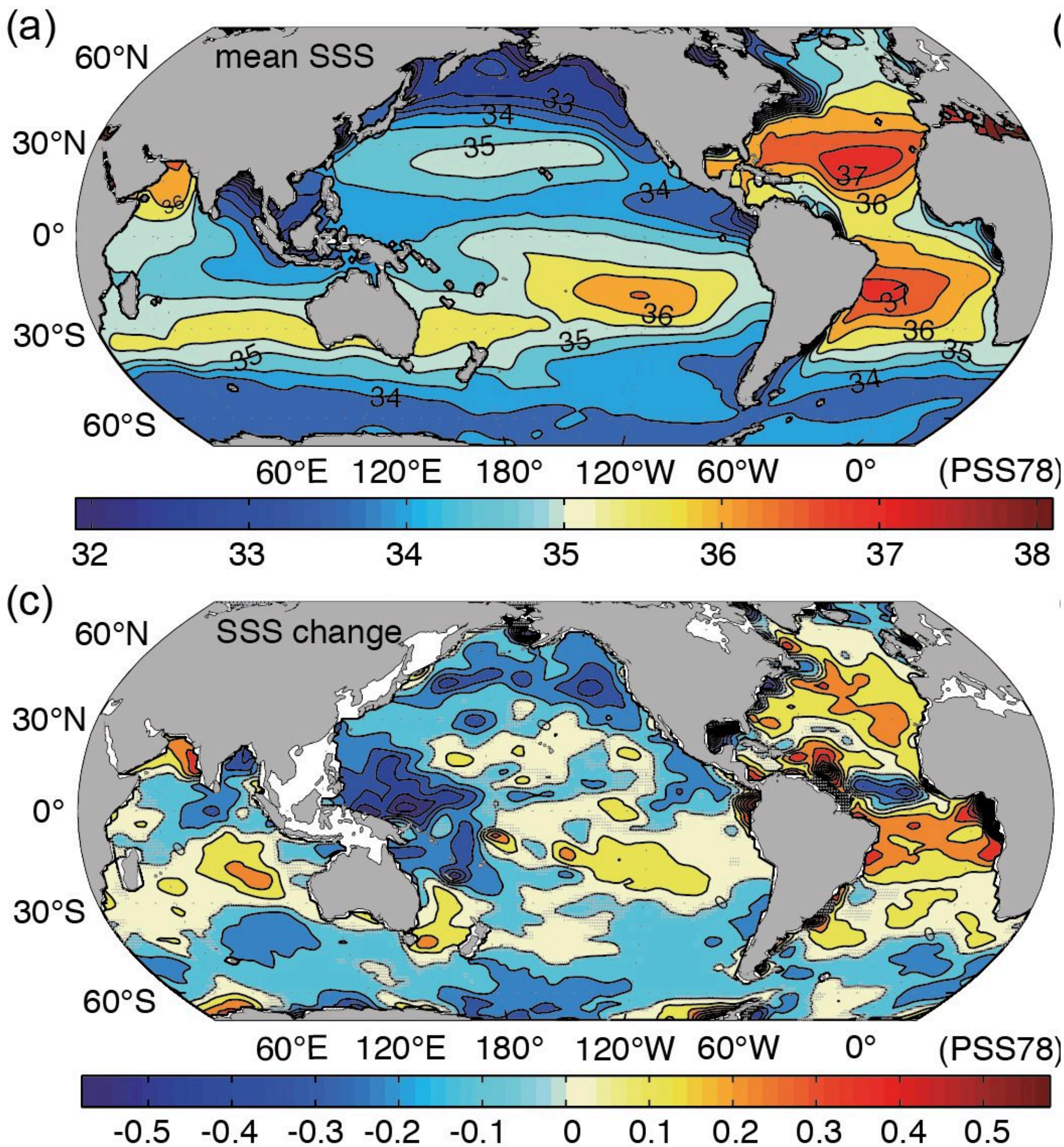


Figure 21: Surface ocean salinity (top) and salinity changes from 1950 to 2008 (bottom) in g/kg. From ipcc.ch.

Warming and freshening of surface waters at high latitudes decreases their density and increases their buoyancy. This reduces deep water formation and the meridional overturning circulation. A reduction of the Atlantic meridional overturning circulation, which was initially predicted by climate models in the 1990's (Manabe and Stouffer, 1993), is now observed in measurements from the subtropical Atlantic (Smeed et al., 2014). However, due to the relatively short period of the available data (2004 to the present) it is currently not clear how much of the observed reduction is caused by human greenhouse gas emissions and how much is due to natural variability. A long-term decline of the Atlantic meridional overturning circulation has been suggested by Rahmstorf et al. (2015) to cause the reduced warming over the subpolar North Atlantic in surface temperature observations (Chapter 2, Fig. 2). Here is a nice interactive article about the topic.

Questions

- Why is it warmer at the equator than at the poles?
- Without heat transport in the atmosphere and oceans, how would temperatures be different at the equator and the poles?
- What determines the surface air pressure?
- What is the Hadley cell?
- How does the Hadley circulation influence precipitation patterns?
- What is the Intertropical Convergence Zone?
- From which direction blows the wind at the surface in the tropics, from which direction does it blow at mid-latitudes?
- In which direction does the Coriolis force acts in the northern and in the southern hemisphere?
- How does the Coriolis force impact the jet stream, the trade winds, and the westerly winds at mid-latitudes?
- What is latent heat of vaporization/condensation?
- How much more heat is required to vaporize water than to heat if from the melting to the boiling point?
- What is the lapse rate?
- How does vertical water vapor transport impact the lapse rate?
- When does evaporation occur?
- When does condensation occur?
- How much larger is the heat capacity of one cubic meter of water than one cubic meter of air?
- How does the difference in heat capacities of air and water impact climate variations?
- Which component of the climate system absorbs most of the energy that currently accumulates on Earth?
- Use the numbers in Fig. (14) to calculate the residence time of water in the atmosphere?
- Use the numbers in Fig. (14) to calculate the residence time of water in the ocean?
- In which direction do surface waters flow in the tropics, in which direction do they flow at mid-latitudes?
- What are the subtropical gyres, and what forces them.
- What is the strongest ocean current in the world?
- What is convergence, what is divergence? What does it imply for vertical flow (upwelling/downwelling)? Name one region each where surface ocean waters converge/diverge?
- What is stratification?
- What is the thermocline?
- What determines sea water density?
- What is the thermohaline circulation?

- Where do surface waters sink into the ocean's interior?
- Where do they upwell back to the surface?
- Why is the Atlantic saltier than the Pacific?
- Where is the ocean warming the most?
- Is it possible that the warming observed in the atmosphere during the past 50 years was caused by changes in ocean circulation? Why?
- Where is the surface ocean becoming saltier, where fresher? How is the pattern of salinification/ freshening related to the salinity of the surface ocean?
- How are changes in surface salinity related to changes in the atmospheric hydrological cycle?

Videos

[Lecture: Atmospheric Circulation]
[Lecture: Hydrological Cycle]
[Lecture: Ocean Circulation]
[YouTube: How Ocean Circulation Works] from It's Okay To Be Smart

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7. Models

Climate models are tools used in climate research. They are attempts to synthesize our theoretical and empirical knowledge of the climate system in a computer code. This chapter describes how climate models are constructed, and how they are evaluated, and it discusses some applications.

a) Construction

Climate models solve budget equations numerically on a computer. The equations are based on the **conservation of energy, momentum, and mass** (air, water, carbon, and other relevant elements, substances, and tracers). Typically they are solved in separate **boxes** that represent specific regions of Earth's climate system components (Fig. 1). Along their boundaries the boxes exchange energy, momentum and mass. Exchange with the flow of water or air from one box to another is called advection. Prognostic variables such as temperature, specific humidity in the atmosphere, or salinity in the ocean, and three velocity components (zonal, meridional, and vertical) are calculated in each box. The momentum equations, which are used to calculate the velocities, are based on Newton's laws of motion and they include effects of the rotating Earth such as the Coriolis force. The temperature equations are based on the laws of thermodynamics. Thus, climate models represent the fundamental laws of physics as applied to Earth's climate system.

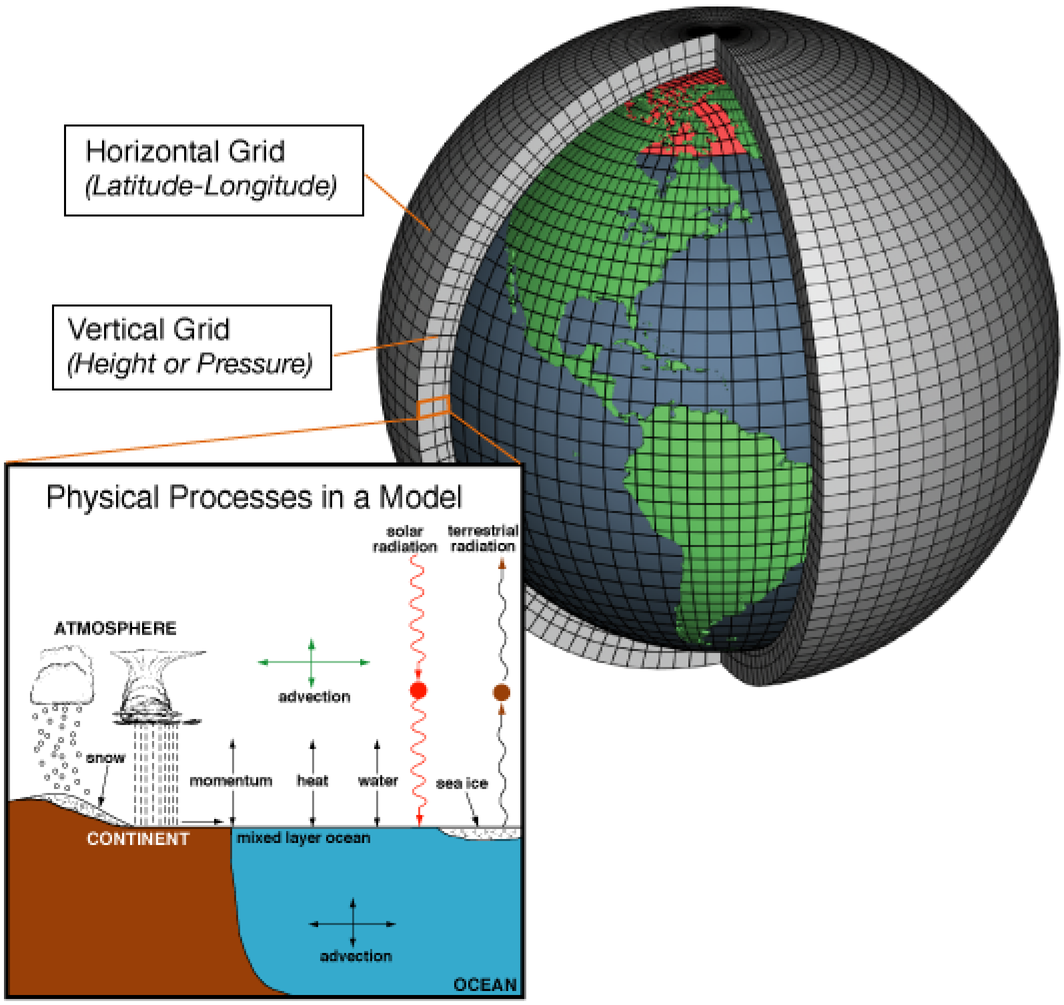


Figure 1: Schematic of a three-dimensional coupled climate model. From wikipedia.org.

The evolution of the prognostic variables in the interior boxes are solved one **time step** at a time (see chapter 4, equation B1.4). After the prognostic variables have been updated the fluxes between boxes (I and O) are calculated, which are used for the next time step. Then the prognostic variables are updated again using the fluxes, and so on. This procedure is called forward modeling because all model variables at the next time step are calculated only from the model variables at the previous time step and the **boundary conditions** without the use of observations. Boundary conditions such as the incident solar radiation at the top-of-the-atmosphere or concentrations of greenhouse gases are usually required as input to the calculations. These are also called radiative forcings. Other boundary conditions need to be applied at the lower boundary: the topography (mountains) and bathymetry (sea floor). To start a forward model simulation **initial conditions** need also to be provided. Those can be taken from observations or idealized distributions.

Climate models range from the simplest zero-dimensional (0D) Energy Balance Model (EBM) discussed in chapter 4 to the most complex three-dimensional General Circulation Models

(GCMs). The range of models ordered with respect to complexity is called the **hierarchy** of climate models. The 0D-EBM can be expanded by solving the energy budget equation separately at different latitudinal bands. This is called the one-dimensional EBM (1D-EBM). The 1D-EBM is still vertically averaged but it includes energy exchange between latitudinal bands as discussed in chapter 6. Meridional energy transport in 1D-EBMs is typically treated as a diffusive process proportional to the temperature gradient, such that heat flows from warm to cold regions. 1D-EBMs typically treat the ocean and atmosphere as one box, so that one cannot distinguish between heat transport in the ocean and atmosphere.

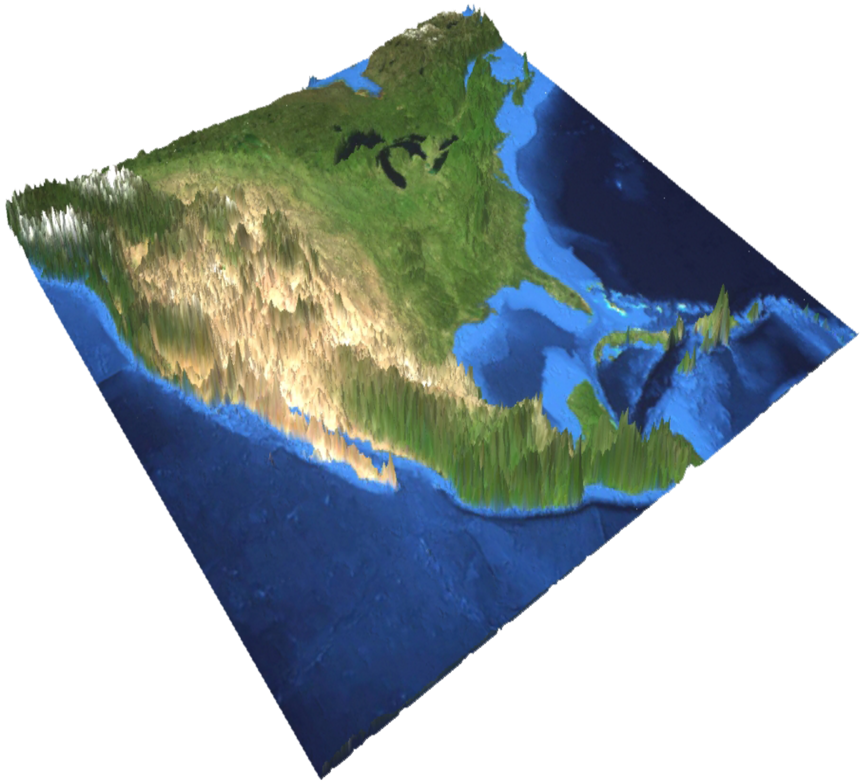
One-dimensional models are also used for vertical energy transfer in the atmosphere. These are called radiative-convective models and they work similar to the models used to produce Fig. (6) in chapter 4. Radiative-convective models themselves range in complexity from line-by-line models, which calculate radiative transfer in the atmosphere at individual wavelengths, to models that average over a range of frequencies (band models). Line-by-line models are computationally expensive and cannot be used in three-dimensional GCMs, which use band models. Band models are calibrated and tested by comparison to line-by-line models. As discussed in chapter 4, radiative fluxes alone would cause a much warmer surface and much colder upper troposphere than observed. Therefore, radiative-convective models include convection, mostly by limiting the lapse rate to the observed or moist adiabatic rate.

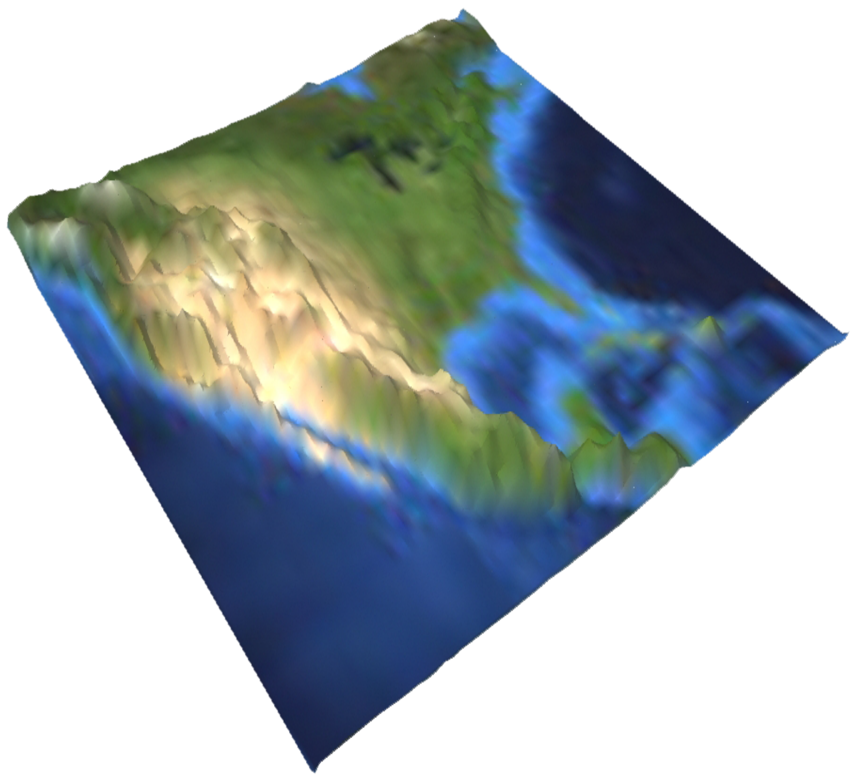
Intermediate complexity models consist of 2D-EBMs, which are still vertically averaged but include zonal transport of energy and moisture (in this case they are called Energy-Moisture-Balance Models or EMBMs), zonally-averaged ocean models coupled to a 1D-EBM, and zonally averaged dynamical atmospheric models (resolving the Hadley circulation). Intermediate complexity models also often include biogeochemistry and land ice components. They are computationally relatively inexpensive and can be run for many thousands or even millions of years.

The first climate models developed in the 1960s were 1D-EBMs and simple GCMs of the **atmosphere** and **ocean** at very coarse resolution. Initially ocean and atmospheric models were developed separately and only later they were coupled. Coupling involves the exchange of heat and water fluxes and momentum at the surface. Current state-of-the-science coupled, three-dimensional GCMs also include **sea ice** and **land surface** processes such as snow cover, soil moisture and runoff of water through river drainage basins into the ocean. Many models also include dynamic **vegetation** with separate plant functional types such as trees and grasses. However, most current coupled GCMs that are used for future projections do not include interactive **ice sheet** components. This is because ice sheets have long equilibration (response) times of tens of thousands of years and therefore they need to be run for a much longer time than the other climate system components, which is currently not possible for most climate models.

The deep ocean has an equilibration time of about a thousand years. For future projections people are mostly interested in the next hundred or perhaps a few hundreds of years. For the most reliable and detailed projections on these timescales global climate modeling groups try to configure their models at the finest possible **resolution**. Currently the typical resolution is a few degrees (~200 km) in the horizontal directions and 20 to 30 vertical levels each in the atmosphere and ocean components. Finer resolution global models are being developed at various climate modeling centers around the world, but currently most climate projections of centennial and longer time scales are based on coarser resolution models (Fig. 2).

Figure 2: Illustrations of land surfaces (green and brown colors) and sea floor (blue) at fine (top; ~10 km) and coarse (botton; ~100 km) resolution. Note that the elevations and depths are strongly exaggerated with respect to horizontal distances. From ucar.edu.





Ice sheet models have finer spatial resolution (10s of kilometers) in order to resolve the narrow and steep ice sheet margin, but they have much larger time steps (1 year) than atmospheric (seconds) and ocean (minutes to hours) models because of the slow ice velocities (10-100 m/yr) compared to velocities of ocean currents (1-10 cm/s) or winds (1-10 m/s). The time step in a model depends on the velocity of the fluid and the grid-box size. The higher the velocity and the smaller the grid-box size the smaller the time step has to be to guarantee numerical stability. Finer resolution models therefore have to use smaller time steps, which is an additional burden on the computational resources. Another obstacle to move to higher resolution is the amount of data that accumulate. The highest resolution ocean model simulations currently require petabytes (10^{15} bytes = 1000 terrabytes) of storage for all the model output. Processing these huge amounts of data is a challenge.

One way to avoid increasing computational resources at higher resolution is to construct **regional climate models** for a specific region of interest, e.g. North America. However, the disadvantage of regional climate models is that boundary conditions at the margins of the model domain have to be prescribed. The resulting solution in the interior depends strongly on those boundary conditions, which often are taken from global climate models. Therefore, any bias from the global climate model at the boundary would be propagated by the regional model into the interior of the model domain. However, in the interior the regional climate model can account for details e.g. of the topography, that a global model cannot. Thus, although not a silver bullet, regional climate models are useful for simulating climate in more spatial detail than possible with global models.

Typically the resolution and grids of the atmospheric and ocean components are different. Therefore, surface fluxes and variables needed to calculate the fluxes need to be mapped from one grid to the other. This is often accomplished by a coupler, which is software that does interpolation, extrapolation, and conservative mapping. All calculations need to be numerically sound such that energy, water, and other properties are conserved. However, numerical schemes, e.g. for the transport from one box to the next, are associated with errors and artifacts.

Models that include **biogeochemistry**, such as the **carbon cycle**, and/or ice sheets are called **Earth System Models**. Earth System Models calculate atmospheric CO_2 concentrations interactively based on changes in land and ocean carbon stocks. They can be forced directly with emissions of anthropogenic carbon, whereas models without carbon cycles need to be forced with prescribed atmospheric CO_2 concentrations.

Due to the limited resolution of the models, processes at spatial scales below the grid box size cannot be directly simulated. For example, individual clouds or convective updrafts in the atmosphere are often only a few tens or hundred meters in size and therefore cannot be resolved in global atmospheric models. Similarly, turbulence and eddies in the ocean, which are important for the transport of heat and other properties, cannot be resolved by global ocean models. These processes must be parameterized. A **parameterization** is a mathematical description of the process that depends on the resolved variables, e.g. the mean temperature in the grid box, and one or more parameters. A simple example of a parameterization is the meridional heat flux $F_m = -K\partial T/\partial y$ in a 1D-EBM, which can be parameterized as a diffusive process, where $K>0$ is the diffusivity, $\partial T/\partial y$ is the meridional temperature gradient and y represents latitude. This parameterization transports heat down-gradient (note the minus sign), which means from warmer to colder regions. In this case the parameter K can be determined from observations of meridional heat flux (Chapter 6, Fig. 4) and $\partial T/\partial y$ (Chapter 6, Fig. 1). Parameterizations can be derived from empirical relationships based on detailed measurements or high-resolution model results. The parameter values are usually not precisely known but they influence the results of the climate model. Therefore, parameterizations are a source of error and uncertainty in climate models. Parameters in the cloud parameterization of a model, for example, will impact its cloud feedback and therefore its climate sensitivity.

b) Evaluation

Climate models are evaluated by comparing their output to observations. Fig. 3 shows the multi model mean (the average of all models) from the most recent IPCC report (Flato et al., 2013). The model simulated surface temperature distribution is similar to the observations (Chapter 6, Fig. 1) with warm (20-30°C) temperatures in the tropics and cold (<0°C) temperatures near the poles and at high altitudes (Himalayas). The models also reproduce some of the observed zonal gradients such as the cooler temperatures in the eastern equatorial Pacific compared to the western Pacific warm pool and the warmer temperatures in the northeast Atlantic compared to the northwest Atlantic, which are caused by the upper ocean circulation. However, the models are not perfect as indicated by biases such as too cold temperatures in the northern North Atlantic and too warm temperatures in the southeast Atlantic and Pacific. The warm biases in the southeast Atlantic and Pacific are most likely caused by the coarse resolution ocean models that do not resolve well the narrow upwelling in these regions. A similar bias is seen in the California current in the northeast Pacific. Despite these biases the multi model mean agrees with the observed temperatures to within plus/minus one degree Celsius in most regions. Even the larger regional biases such as those mentioned above are relatively small compared to the ~60°C range of temperature variations on Earth. This indicates that the models reproduce observed surface temperatures relatively well.

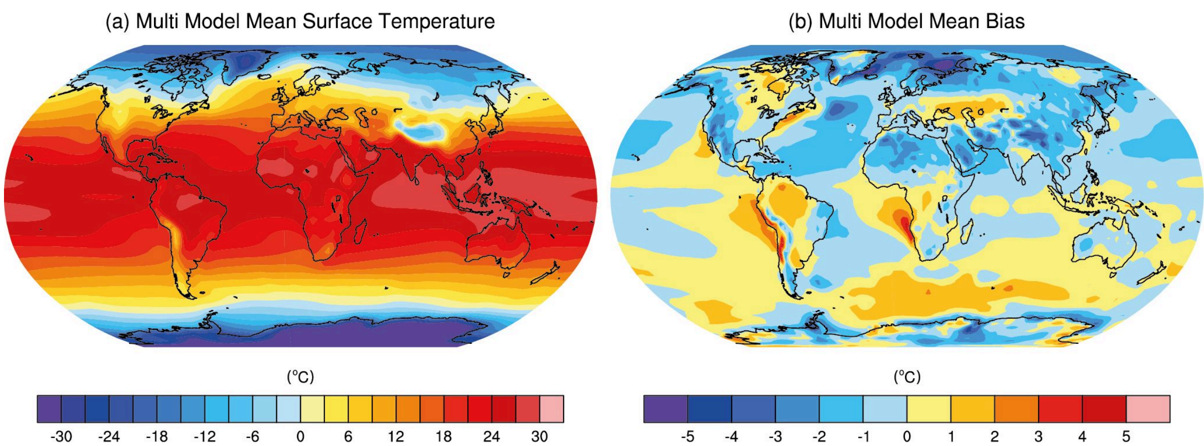


Figure 3: Annual mean surface air temperature distribution from 1980-2005 as simulated by climate models (a) and their bias (b). The bias b = $T_m - T_o$ is the difference between the model simulated temperature T_m and that estimated from observations T_o. From ipcc.ch.

Fig. 4 shows the multi model mean precipitation. The models reproduce the general pattern of the observations (Chapter 6, Fig. 8) such as more precipitation in the tropics and at mid-latitudes compared with less precipitation in the subtropics and at the poles. They also reproduce some of the observed zonal differences such as dryer conditions over the eastern parts of the subtropical ocean basins compared with wetter conditions further west. However, the models also display systematic biases such as the double Intertropical Convergence Zone (ITCZ) over the East Pacific and too dry conditions over the Amazon. The relative errors in precipitation are generally larger than those for temperature. This indicates that the models are better in simulating temperature than precipitation. This may not be surprising given that the simulation of precipitation depends strongly on parameterized processes such as convection and clouds.

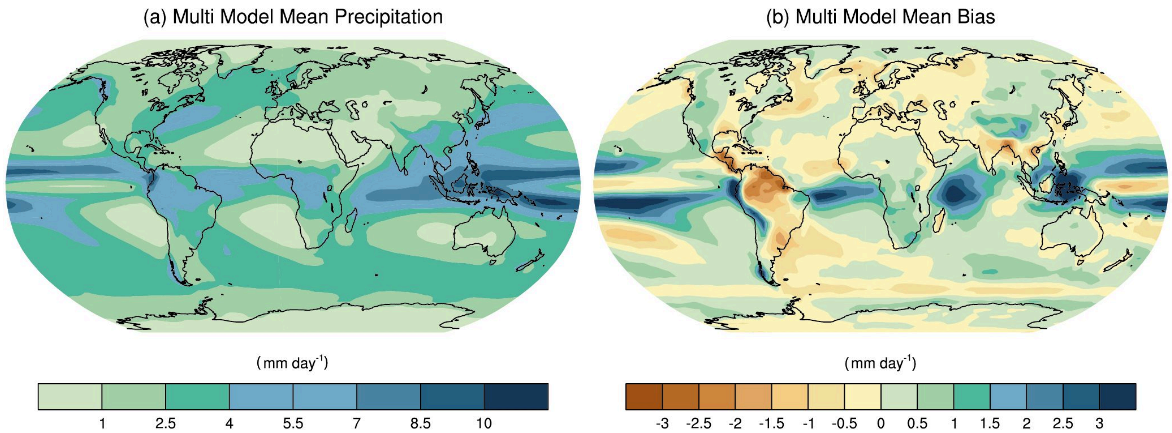


Figure 4: As Fig. 3 but for annual mean precipitation.

Fig. 5 compares correlation coefficients of different variables. It confirms our previous conclusion that the models are better in simulating temperature than precipitation. It also shows that the models have very good skill in simulating Emitted Terrestrial Radiation at the top-of-the-atmosphere, whereas they are less good at simulating clouds. The current generation of climate models (CMIP5) are improved compared with the previous generation (CMIP3) particularly for precipitation. Another interesting feature also apparent in Fig. 5 is that the the multi model mean is almost always in better agreement with the observations than any one particular model. A similar phenomenon, which has been called the wisdom of the crowd, has been noted by Sir Francis Galton (1907), who analyzed villager's guesses for the weight of an ox at an English livestock fair. He found that many guesses where too high or too low, but the mean of all guesses was almost exactly the correct weight of the animal.

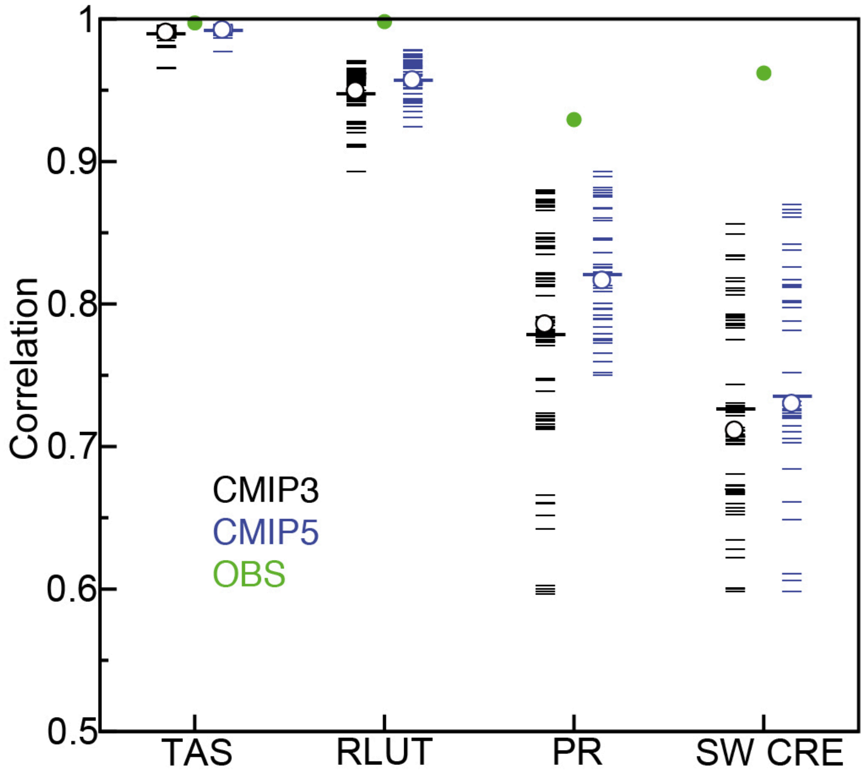


FIgure 5: Correlation coefficient for the annual mean surface air temperature (TAS), outgoing longwave radiation (RLUT; in chapter 6 we called this variable emitted terrestrial radiation from the atmosphere: ETR$_a$*), precipitation (PR), and the shortwave cloud radiative effect (SW CRE). The correlation coefficient* $-1 \leq r \leq 1$ *is a measure of pattern agreement. See this [wikipedia.org] page for a detailed definition of r. If two variables are unrelated* $r = 0$*, whereas if they are perfectly in linear agreement* $r = 1$*. Each short horizontal line is one model. The longer line is the multi model mean and the circles are the median. Results are shown for the current generation of climate models from the Climate Modeling Intercomparison Project 5 (CMIP5) and the previous generation (CMIP3). Green dots show correlation coefficients for alternative observational datasets indicating uncertainties in the observations. They represent the best the models could do given uncertainties in the observations. From [ipcc.ch].*

Fig. 6 shows that most models overestimate temperatures in the thermocline by about 1°C, presumably due to too much vertical diffusion. Again these errors are relatively small given the large range (~20°C) of deep ocean temperature variations. Model errors in salinity are larger near the surface than in the deep ocean. In the southern hemisphere subtropics near surface waters are too fresh, perhaps related to the double ITCZ bias and associated too wet conditions in the atmosphere there (Fig. 5).

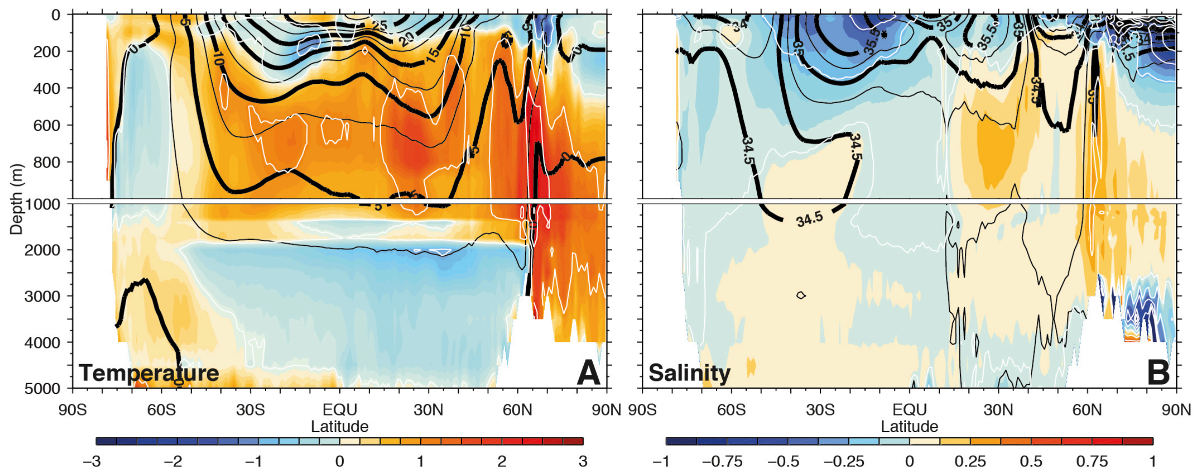


Figure 6: Zonally averaged differences in temperature (A) and salinity (B) between the multi model mean and observations (color contours and white lines). Black lines indicate observed temperature and salinity from the World Ocean Atlas 2009. From ipcc.ch.

All climate models simulate an increasing ocean heat content over the last 40 years, consistent with observations. Some models simulate more and others less heat uptake. The multi model mean, however, is in good agreement with the observations. This indicates that the models are skillful not only in simulating the mean state of ocean temperatures but also its recent temporal evolution.

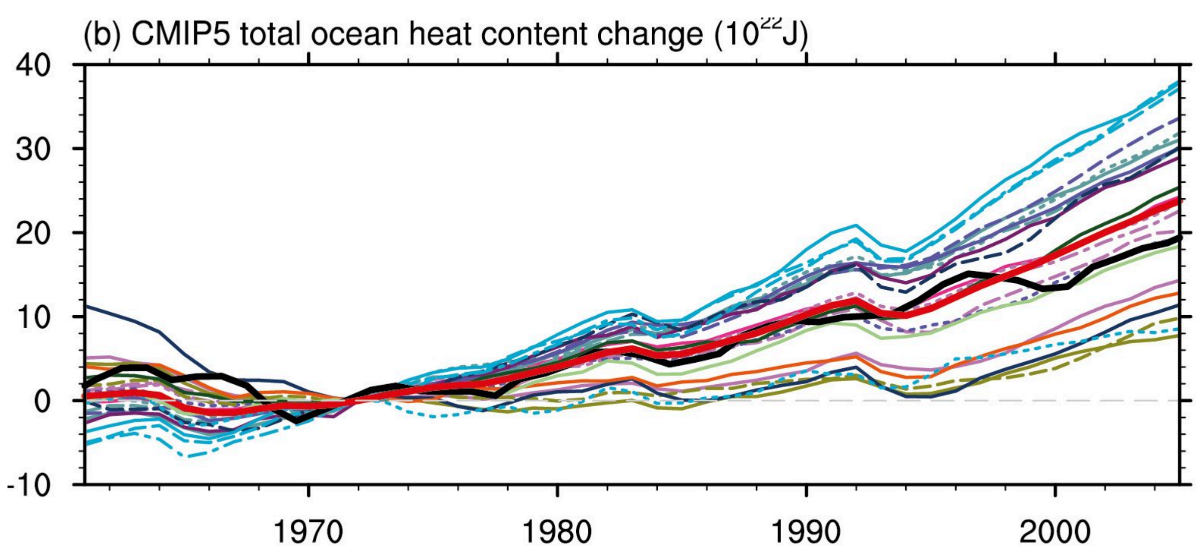


Figure 7: Changes in ocean heat content simulated by the CMIP5 models (thin lines), and their multi model mean (red thick line), compared to observations (black thick line). From ipcc.ch.

Historical and paleoclimate variations are also used to test and evaluate climate models as will be discussed next.

c) Applications

Some of the main applications of climate models are paleoclimate studies, detection and attribution studies, and future projections. Future projections will be discussed in the next chapter.

Paleoclimate model studies are not only useful for a better understanding of past climate changes and their impacts but they can also be used to evaluate the models. For example, model simulations of the Last Glacial Maximum (LGM) are broadly consistent with temperature reconstructions (see Chapter 3) that show global cooling of 4-5°C, polar amplification and larger changes over land than over the ocean. The models reproduce these basic features of the reconstructions, which indicates that they have skill simulating climates different from the present (Masson-Delmotte et al., 2013). On the other hand, there are also differences between model results and observations, e.g. in deep ocean circulation (Muglia et al., 2015), which indicates that model's skills in those aspects are questionable.

Detection and attribution studies attempt to determine which observed climate changes are unusual (detection) and what are its causes (attribution). Climate models driven with only natural forcings show variations from one year to the next due to internal climate variability (e.g. El Niño) and short term cooling in response to large volcanic eruptions (Fig. 8). However, they do not show a long-term warming trend over the past century, in contrast to the observations. This suggests that the global warming observed since about the 1970's is highly unusual and cannot be explained by internal climate variability (as represented in the models) nor by natural drivers. However, if anthropogenic forcings are included the models reproduce very well the observed long term trend. The multi model mean also reproduces well the observed short term coolings associated with large volcanic eruptions of the past 50 years.

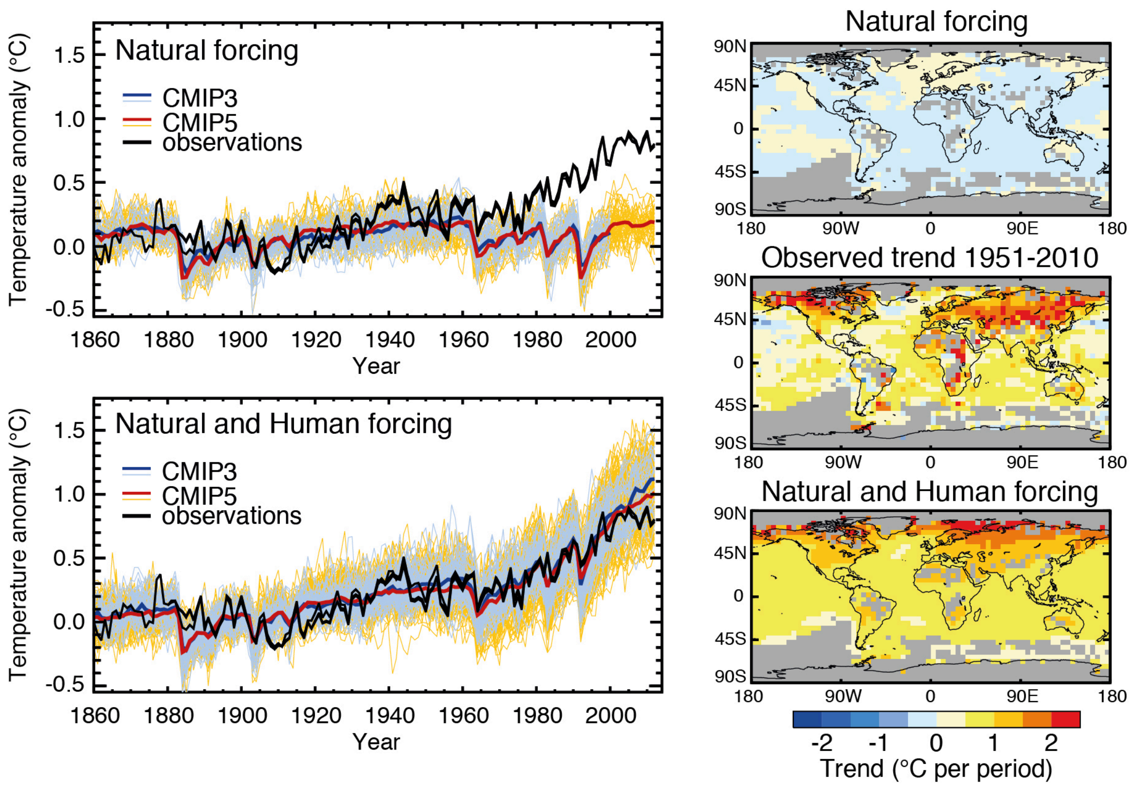


Figure 8: Comparison of climate model simulations with observed surface temperature changes. Left panels show timeseries of globally averaged surface temperatures from the current generation of climate models (CMIP5; yellow thin lines show individual models, red thick line shows the multi model mean) and the previous generation (CMIP3; blue lines, mostly obscured by the overlain yellow lines). Two sets of simulations are show. Top panels show models forced with natural forcings only, bottom panels show models forced with both natural and human forcings. From ipcc.ch. This is a key figure.

Models driven with both natural and anthropogenic forcings reproduce well not only the observed global mean temperature changes but also its spatial distribution such as larger warming at high northern latitudes (polar amplification) and over land (land-sea contrast). These results represent evidence that human activities are the main reason for the observed warming during the past 50-60 years. The IPCC's AR5 concludes that "it is extremely likely that human activities caused more than half of the observed increase in GMST from 1951 to 2010" (Bindhoff et al., 2013), where GMST stands for global mean surface temperature.

Questions

Answer the following questions:

- How do climate models work? Describe the equations that are solved.
- When were the first climate models developed?
- What is the hierarchy of climate models?
- List three different types of climate models of different complexity.
- Which components of the climate system are currently included interactively in most climate models?
- Which components are currently not included in most models?
- What is an Earth System Model?
- What is the typical resolution of a General Circulation Model (include units)?
- How many layers does a typical three-dimensional climate model have in the ocean?
- How many does it have in the atmosphere?
- Are observations used in forward climate model simulations?
- What are some inputs (boundary conditions) required for a climate model simulation? List two.
- How are climate models evaluated?
- Which variable to climate models simulate better: temperature or precipitation?
- What are climate models used for? List two applications.
- How are climate models used to attribute recent climate change?
- What is the result of those attribution studies?

Videos

[Recorded Lecture: Climate Models]

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8. Impacts

Climate change impacts not only physical but also biological and human systems. Possible impacts of future climate change are important considerations for current decision making, but they are associated with uncertainties. In this chapter we will discuss future projections with climate models, the main sources of uncertainty for those projections, and some global scale impacts.

a) Projections

Prediction is difficult, especially about the future[1]. This is also true for projections of future climate change. They rely on assumptions about future radiative forcings such as anthropogenic emissions of greenhouse gases and aerosols, which are unknown. For this reason they are called projections and not predictions.

Climate scientists believe that the best projections consider results from state-of-the-science climate models because they are syntheses of theoretical and empirical knowledge. However, as we have learned in the previous chapters, climate models are imperfect. **Uncertainties** in future projections will therefore arise from both, assumptions about future greenhouse gas emissions and climate model errors.

The most recent IPCC assessment report (AR5) uses future scenarios that specify anthropogenic radiative forcing (Fig. 1; Collins et al., 2013; Stocker et al., 2013). They are called Representative Concentration Pathways (RCPs) and are followed by a number that indicates the radiative forcing around year 2100. The goal is to cover a range of possible futures. Note that the radiative forcing slowly decreases after the year 2050 for scenario RCP2.6. In this case the maximum forcing of ~2.5 Wm^{-2} is reached during the middle of this century. Scenarios RCP4.5 and RCP6.0 continue to increase the radiative forcing during this century and stabilize close to their maximum around year 2100, whereas for scenario RCP8.5 the forcing continues to increase until about year 2200 after which is stabilizes at 12 Wm^{-2}.

1. This quote has been attributed to various people including the Danish physicist Niels Bohr and American writer Mark Twain.

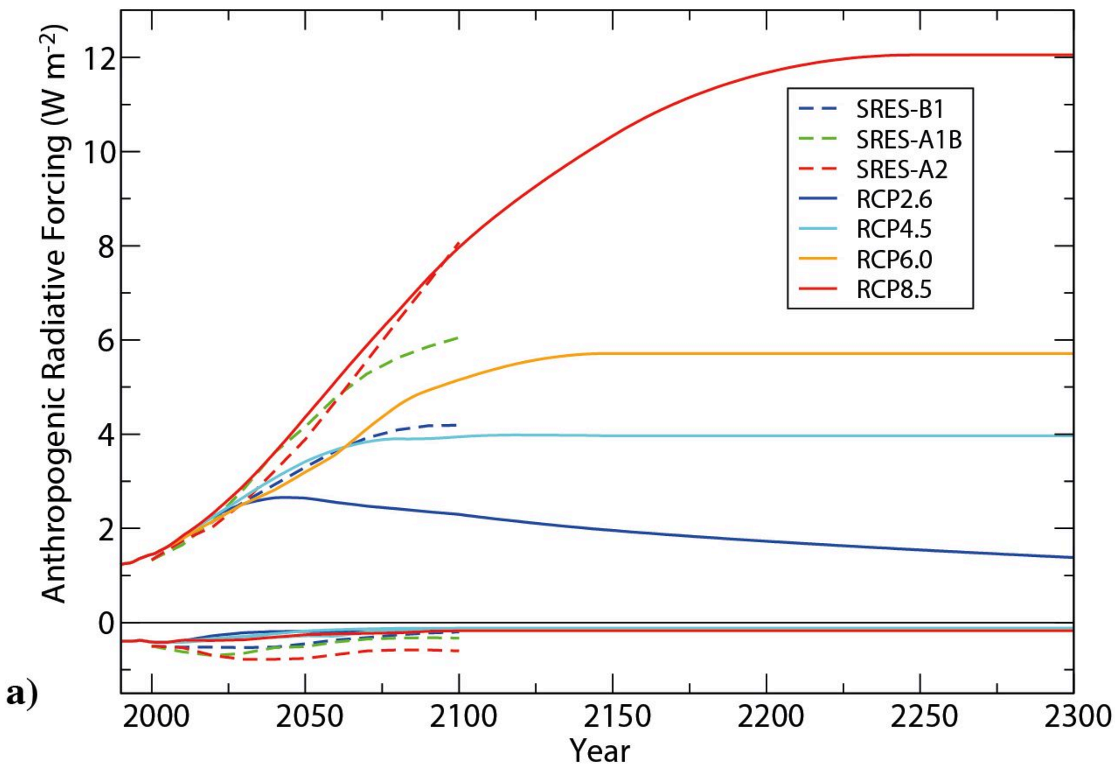


Figure 1: Anthropogenic radiative forcing as a function of time for the current (RCP) and previous (SRES) IPCC scenarios. From ipcc.ch.

The prescribed radiative forcing is then used to calculate corresponding CO_2 concentrations (Fig. 2), e.g. by inverting eq. (2) in chapter 4. Scenario RCP2.6 leads to a stabilization of CO_2 concentrations below 450 ppm during the 21st century, whereas in scenarios RCP4.5 and RCP6.0 CO_2 concentrations continue to increase to about 550 ppm and 650 ppm, respectively, at year 2100. In scenario RCP8.5 CO_2 concentrations increase to about 900 ppm at year 2100. For the upcoming AR6 different scenarios, called Shared Socioeconomic Pathways or SSPs, have been developed.

Carbon cycle models are used to calculate carbon emissions consistent with those CO_2 concentration pathways. Due to uncertainties in the models a range of carbon emission scenarios emerges (shaded bands in Fig. 2). Scenario RCP2.6 implies strong reductions in carbon emissions after about year 2020, whereas the business-as-usual scenario RCP8.5 assumes continued increases in emissions until the year 2100 (Fig. 2). The two intermediate scenarios RCP4.5 and RCP6.0 assume continued increases in emissions until about 2050 or 2080 followed by emissions reductions.

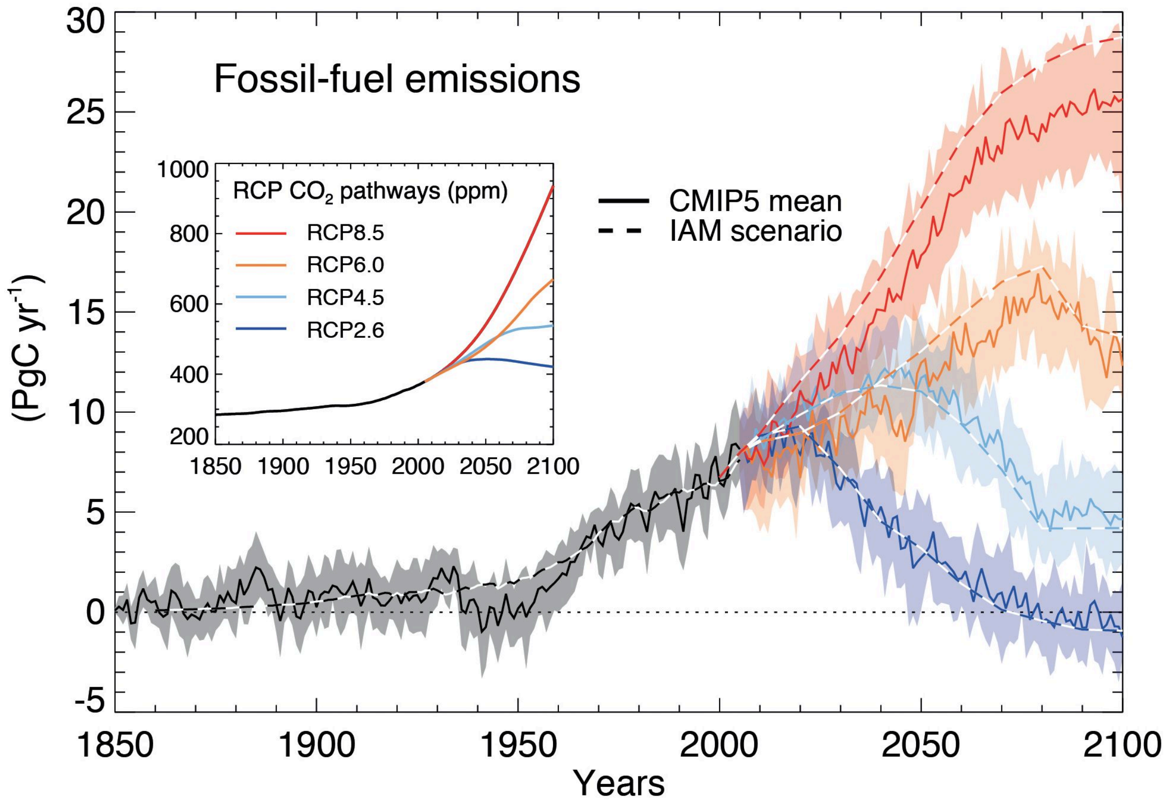


Figure 2: Evolution of CO_2 emissions and concentrations (inset) for RCP scenarios. Emissions are calculated from simple carbon cycle box models given the CO_2 concentrations. Uncertainties in these calculations are indicated by the shading and the two different lines labeled CMIP5 mean and IAM scenario. From ipcc.ch.

The CO_2 concentration pathways are used as input to comprehensive climate models, which project a range of global temperature responses (Fig. 3). For scenario RCP2.6 the models project further warming of less than 1°C above current levels by year 2050 and subsequent slow cooling. Scenarios RCP4.5 and RCP6.0 result in additional warming of about 1.5 to 2°C until year 2100, whereas for the business-as-usual scenario RCP8.5 temperatures increase by 4°C by the year 2100 and by 8°C by year 2300. The latter corresponds to twice the temperature difference between the Last Glacial Maximum and the pre-industrial (see chapter 3). Note that the uncertainty as indicated by the shading is much larger for the high emission scenario RCP8.5 compared with the intermediate and low scenarios.

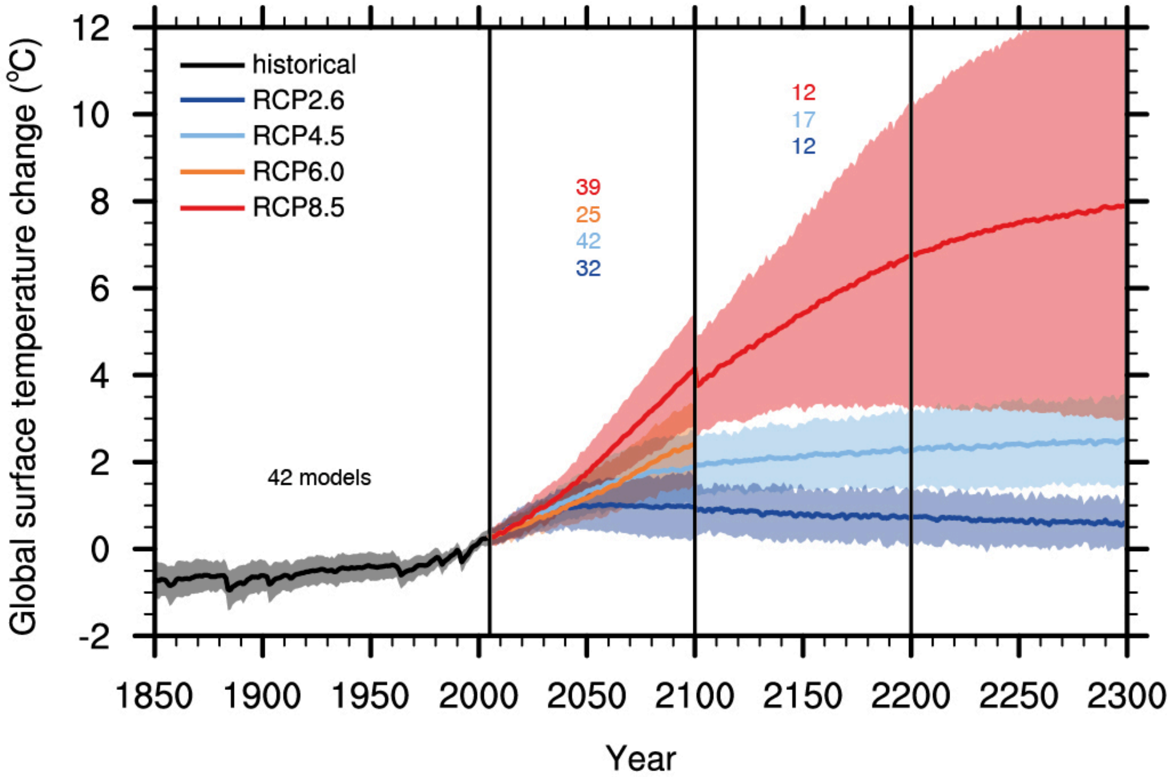


Figure 3. Global average surface temperature projections relative to 1986-2005. Shading indicates the 5 and 95% range and the line is the multi model mean. The black line with the grey shading represents historical simulations including variations in volcanic forcing. Since volcanic forcing is impossible to predict for the future, changes in natural radiative forcings are not considered in the projections. The number of climate models used is indicated. Not all models have simulated the full period until year 2300, which leads to a jump in the red line at year 2100. From ipcc.ch.

The climate models project more warming over land than over the ocean and more warming at high latitudes, especially in the Arctic (Fig. 4). This pattern is similar for all scenarios and time periods. However, the absolute numbers are smaller for lower emission scenarios than for the business-as-usual scenario RCP8.5 shown in Fig. 4. We have encountered these patterns of land-sea contrast and polar amplification before in the observations of historical temperature changes discussed in chapter 2, in reconstructed temperatures from the Last Glacial Maximum discussed in chapter 3, and in our discussions of detection and attribution studies in chapter 7. We conclude that these patterns are robust features of the climate system that are relatively well understood and reproduced in models. Warming in the Arctic is more than twice as much as the global average, which is dominated by the oceans. The IPCC concludes that, on average, warming over land will be 1.4 to 1.7 times larger than warming over the ocean (Collins et al., 2013).

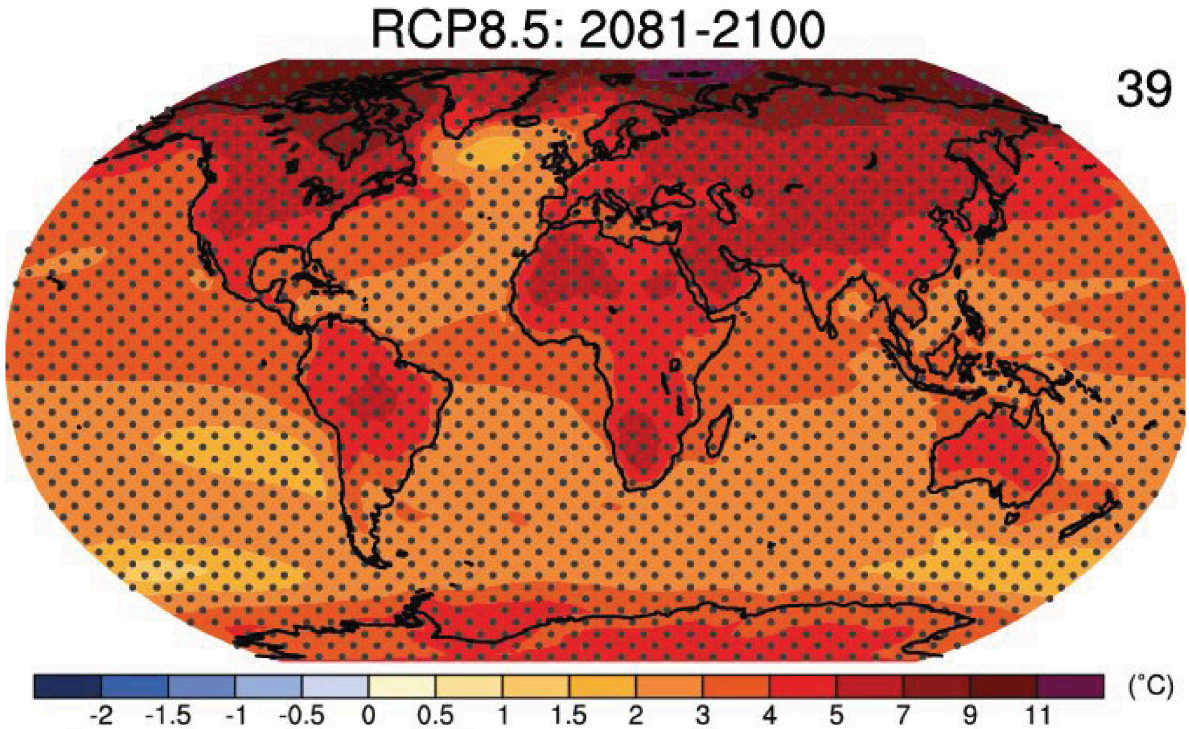


Figure 4: Surface temperature changes projected for the business-as-usual scenario RCP8.5 at the end of the 21st century. Stippling indicates regions of highly significant changes, where the multi model mean is larger than two standard deviations of modeled internal variability and where at least 90% of models agree on the sign of the change. Click on the map to see projections for other scenarios and times. From ipcc.ch.

Explore Projections

Explore projections of future climate using this website:

- Select a country, variable and model of your choice.
- How does the model simulation of the seasonal cycle (lower left plot) for the 1980-2004 period compare with the observations?
- Try a different model and then the model mean. Which agrees better with the observations?
- Compare simulations of temperature and precipitation with observations. Which fits better?
- Now compare the model simulation of the future (2050-2074) with the recent past (1980-2004). Describe your observations.

Another good website with climate model projections but also other data is climexp.knmi.nl.

Fig. (5) shows that models project more warming in the upper tropical troposphere than at the surface. Increased water vapor content in the warmer atmosphere (see discussions in Chapter 4 of the Clausius-Clapeyron relation and the lapse rate feedback, and in Chapter 6 of the hydrological cycle) causes more upward moisture transport in the ascending part of the Hadley cell. Enhanced condensation and latent heat release leads to intensified warming there.

Models also project significant cooling in the stratosphere. Since the stratosphere is stably stratified (there is no convection) its energy balance is determined by local heating from the **absorption** of solar radiation by ozone and local cooling from the emission of infrared radiation by CO_2 (Manabe and Strickler, 1964). Thus, higher CO_2 concentrations in the stratosphere will lead to more emission of infrared radiation and cooling there. Stratospheric cooling is observed in contemporary measurements, consistent with the cooling effect of CO_2. However, the current cooling is also caused by decreases in stratospheric ozone caused by human emissions of Chlorofluorocarbons (CFCs). Nevertheless, the projected stratospheric cooling for the future will be dominated by CO_2 increases since CFC emissions have decreased and the ozone hole is expected to heal over the next decades.

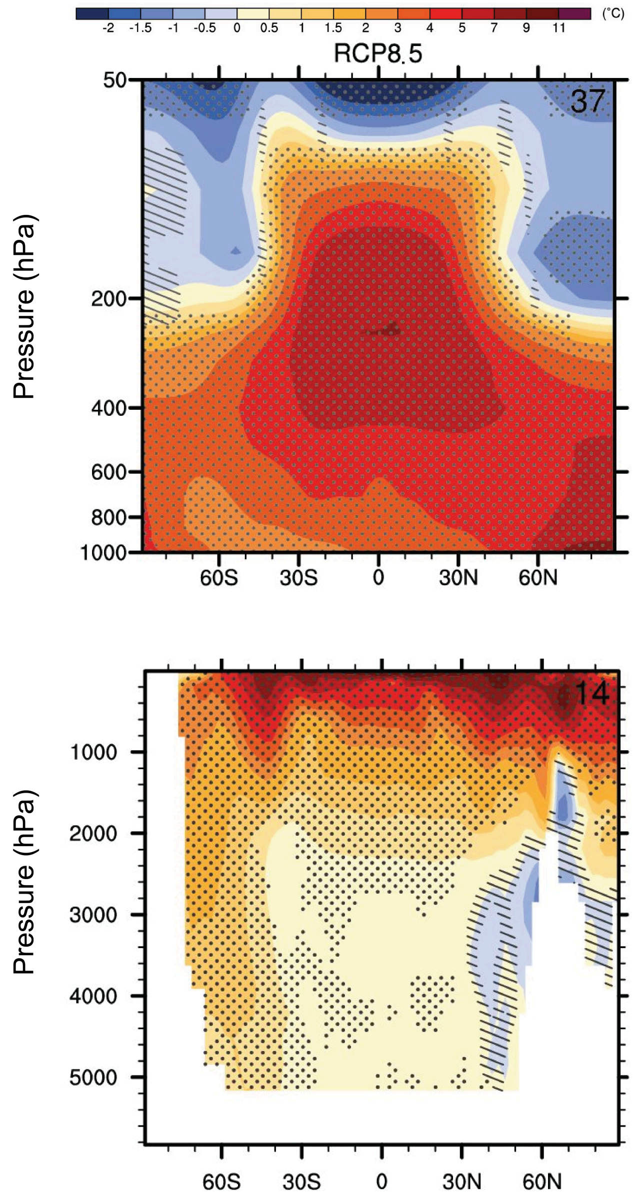


Figure 5: Zonally averaged temperature changes projected for the end of the 21st century in the atmosphere (top) and ocean (bottom) as a function of latitude and pressure (height, top) and depth (bottom) for scenario RCP8.5. Stippling indicates significance as in Fig. 4. Hatching indicates where changes are not significant with respect to 2 sigma variations of internal variability. Click on the ocean panel to see results for other scenarios. From ipcc.ch.

The observed warming of the upper ocean is projected to continue and penetrate deeper into the ocean interior particularly in regions of deep water formation such as the Southern Ocean (Fig. 5). More warming of the surface will increase the stratification of the ocean and reduce vertical mixing, with implications for the ocean's meridional overturning circulation and nutrient delivery to the euphotic zone, both of which will likely continue to decrease.

Warming of the surface is projected to increase evaporation everywhere except in the northern North Atlantic and Southern Ocean where the warming is smallest and over some land areas where precipitation is projected to decrease (Fig. 6). Precipitation is generally projected to increase in regions that are already wet like the tropics and at mid- to high-latitudes, and decrease in regions that are already dry like the subtropics. Relative humidity is generally projected to decrease over land areas and increase over the oceans. The net surface freshwater loss, which is determined by evaporation minus precipitation (E − P), is projected to increase in the subtropics, whereas the tropics and mid- to high latitudes will gain water. Runoff is projected to increase in most regions except around the Mediterranean, the Southwest US and Mexico, southwest Africa, and parts of South America. Soil moisture, on the other hand, is projected to decrease almost everywhere except in parts of North Africa, Asia, and South America, where precipitation increases substantially.

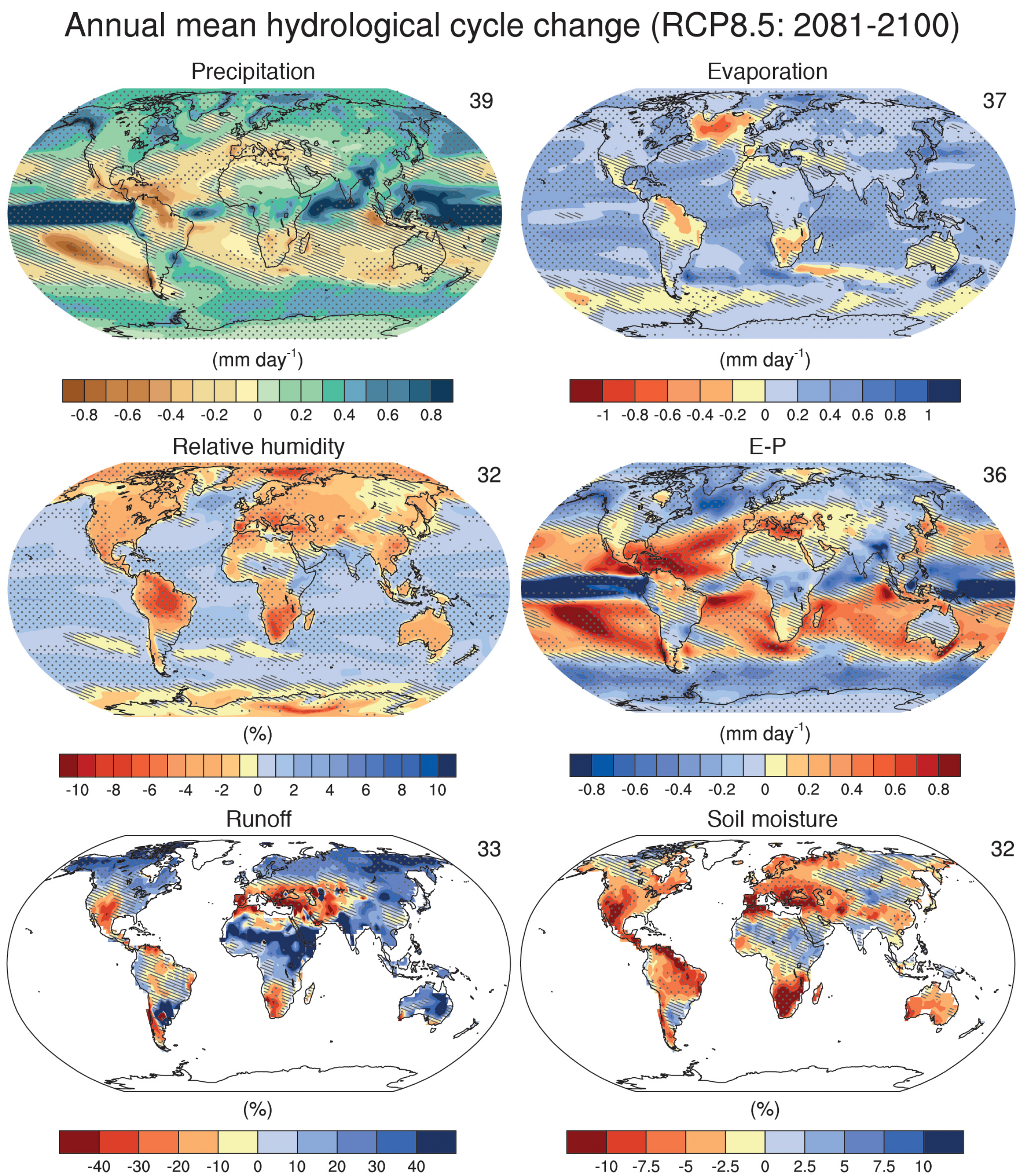


Figure 6: As Fig. 4, but for changes in the hydrological cycle. From ipcc.ch.

Projections suggest that Arctic sea ice will continue to decline in the future particularly at the end of summer (Fig. 7). However, much larger declines are projected for the high emission scenario (RCP8.5), for which all models project an almost completely ice free Arctic ocean, compared to the low emission scenario (RCP2.6), for which all models project a substantial remaining sea ice cover.

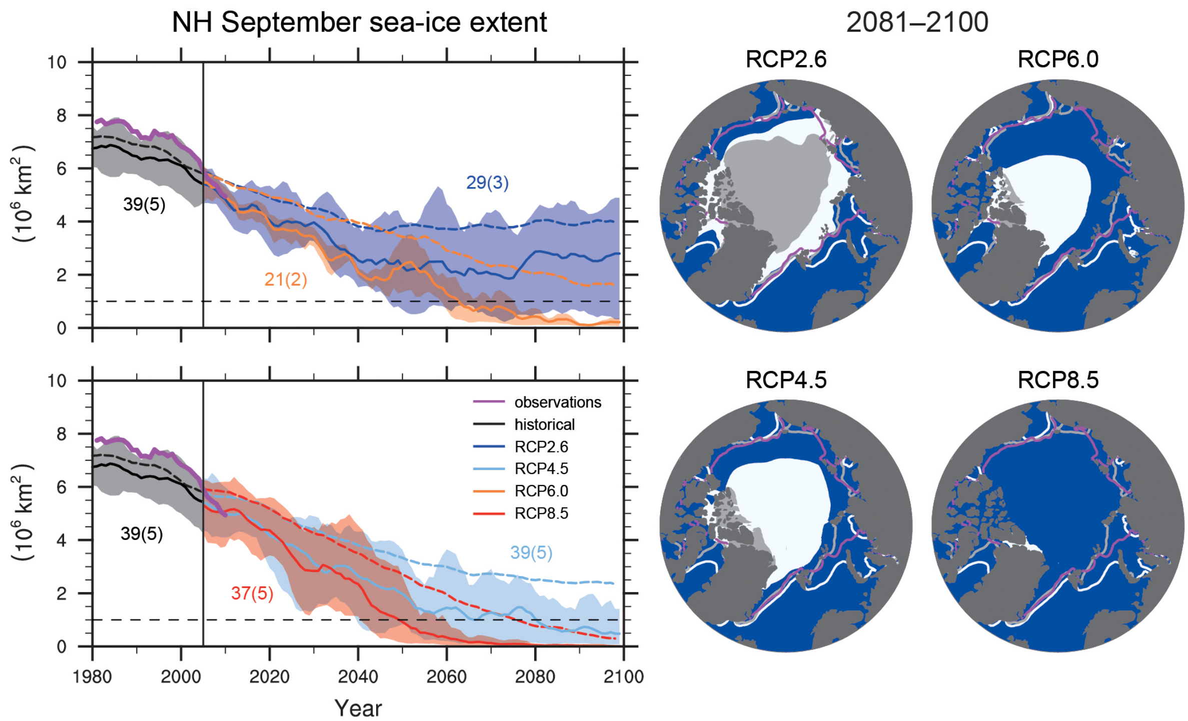


Figure 7: Projections of Arctic sea ice in September. Results from all models are shown as dashed lines in the time series panels on the left and as white shading in the maps on the right. Results from a subset of models that agree best with observed sea ice (numbers in brackets) are shown as solid lines and grey shading. From ipcc.ch.

Mountain glaciers are also projected to continue to melt, as well as the Greenland and Antarctic ice sheets. Due to this input of mass together with the expansion of the ocean from warming **sea levels** are projected to increase by 40 to 70 cm by the year 2100 (Fig. 8). Note that sea level projections are very similar for all scenarios until mid-century. This indicates the **commitment** to the near future sea level rise from past carbon emissions. In other words, considerable climate change and the associated impacts are already baked into the system and cannot be avoided. It implies the **necessity to adapt** to these unavoidable impacts.

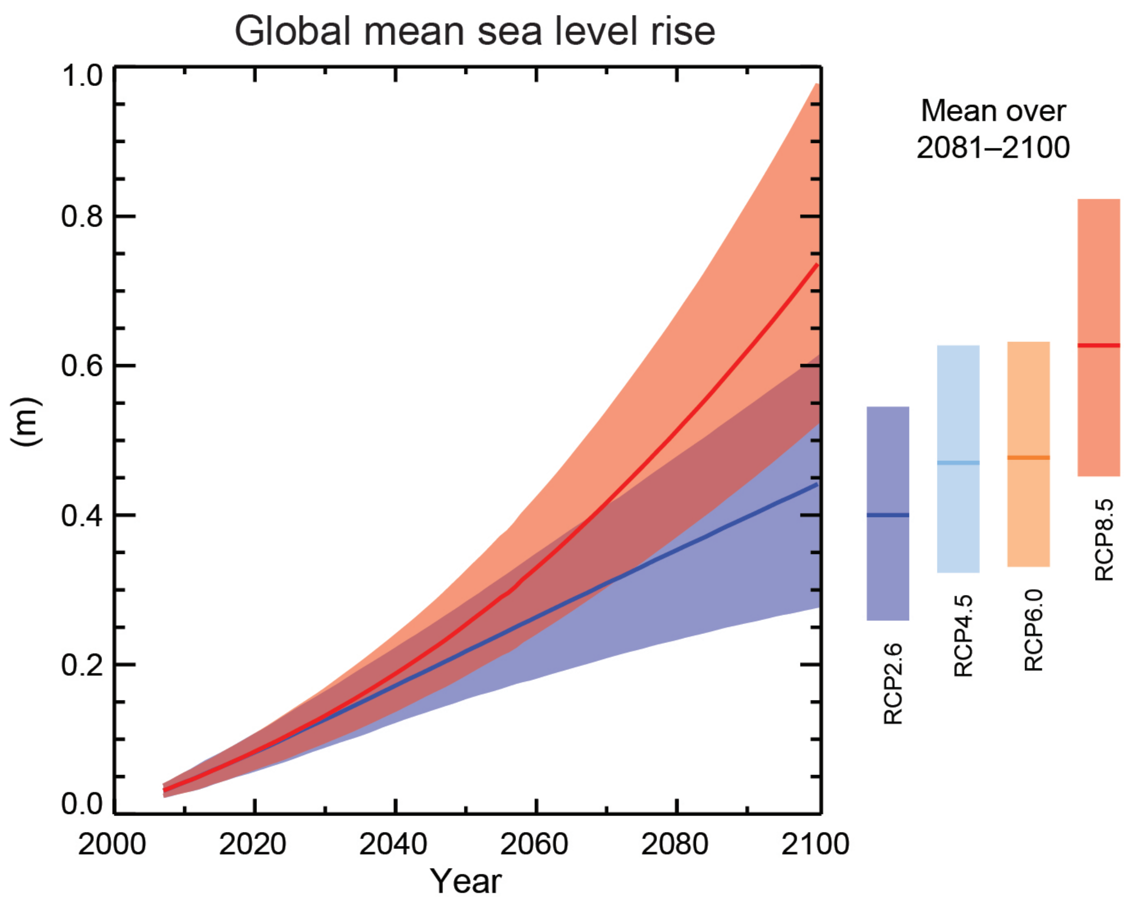


Figure 8: Global sea level rise projected for the 21st century. The blue and red curves show projections for scenarios RCP2.6 and RCP8.5, respectively. The shading indicates the likely (66-100%) uncertainty range. Bars on the right show the 2081-2000 averages for four scenarios. From ipcc.ch.

Increasing stratification from warming and freshening of North Atlantic surface waters will lead to a slowdown of the Atlantic meridional overturning circulation, by about 10% for each degree Celsius global mean warming (Bakker et al., 2016). Many other impacts also depend approximately linearly on global mean temperature and global mean temperature can be well approximated by cumulative carbon emissions (Matthews et al., 2009) such that 500 Gt lead to approximately 1°C of warming (Fig. 9). Since we've already emitted about 500 Gt, this means that if we want to stay below 2°C, the goal of the Paris Agreement, we have only another 500 Gt of carbon to emit.

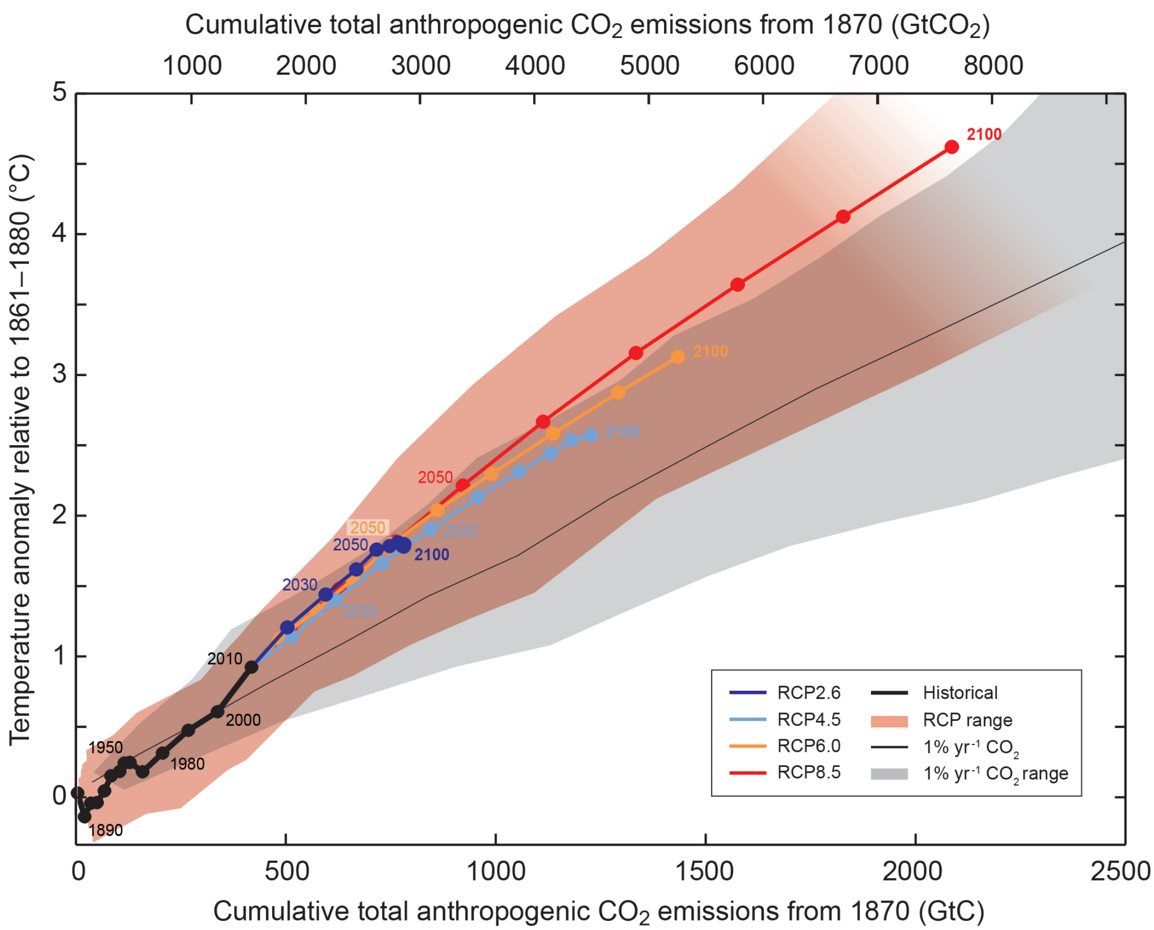


Figure 9: Global mean surface temperature change as a function of the cumulative anthropogenic carbon emissions. Cumulative carbon emissions are the area under the curves in Fig. 2. Colored lines show multi-model means from the RCP scenarios and colored plume indicates uncertainties. Results from exponential CO_2 increase experiments at a rate of 1% per year are shown as the black line and grey shading. RCP scenarios lead to somewhat larger warming because they include additional non-CO_2 forcings. From ipcc.ch.

Box 1: Tipping Points

A tipping point is a threshold that when crossed will lead to an irreversible transition into a different state. Examples for tipping points are

- melting of the Greenland ice sheet,
- melting of the West Antarctic ice sheet,
- collapse of the Atlantic meridional overturning circulation,
- ecosystem shifts,
- species extinctions.

Ice sheets can have tipping points due to positive feedbacks, such as the *ice sheet elevation – mass balance feedback* (Weertman 1961). This feedback can be understood by considering an idealized ice sheet on a flat bed as illustrated in Fig. B1.1 below. An ice sheet gains mass through accumulation of snow in the upper central parts, whereas it looses ice through melting along the lower-lying margins. For simplicity we assume in the following that the accumulation and melt rates are spatially constant. Due to gravity and the resulting horizontal pressure gradient the ice flows slowly from the center towards the margins. Moving upward on a glacier or ice sheet one starts at the margin to walk on ice. At some point the ice turns into snow, which covers the interior of the sheet all the way to the summit. This is the so-called snow line or equilibrium line. It indicates the transition from the melt-zone to the accumulation zone. If the mean height of the ice sheet $\overline{H}$ is larger than the height of the equilibrium line H_E the ice sheet is in a stable equilibrium. However, if warming temperatures rise the equilibrium line to $\overline{H}$ the ablation area increases and the ice sheet looses more mass than it gains. It shrinks and its height drops, which leads to even more melting. This feedback can lead to a rapid and irreversible melting of an entire ice sheet.

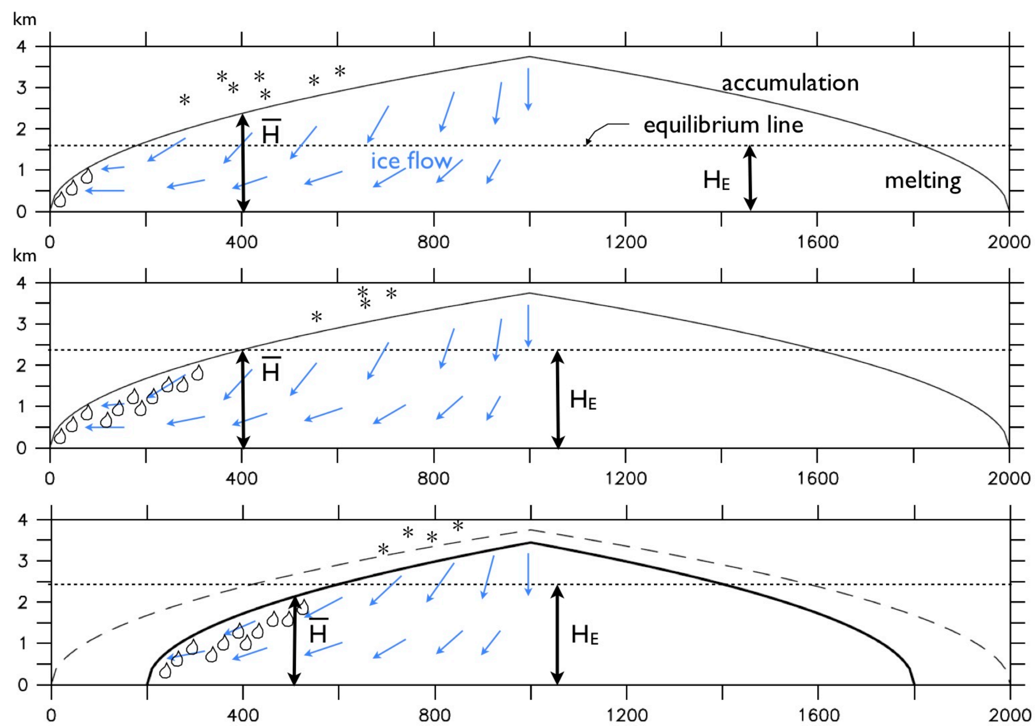


Figure B1.1: Schematic section across an idealized ice sheet. The horizontal axis represents the distance in km from its left margin. The vertical axis represents elevation above the flat base.

As a result of this feedback ice sheets exhibit hysteresis behavior as illustrated schematically in Fig. (B1.2). If the height of the equilibrium line is between zero and $\overline{H}$ two states are possible. One without ice and one with an ice sheet. Assume we're in a state without ice on the lower branch of the hysteresis curve. If the climate cools and the equilibrium line drops below zero ice will grow and we will switch to the state with an ice sheet.

If the climate now warms again the ice sheet will stay in place until the height of the equilibrium line raises above $\overline{H}$. At this point the ice sheet will melt. So, for equilibrium line heights between zero and $\overline{H}$ two states are possible and which state the system is in depends on its history. This is a typical property of hysteresis as observed e.g. in magnetism. It also can lead to rapid changes in state as illustrated by the vertical arrows. Due to its latitude being further poleward and its height being higher it is thought that the East Antarctic ice sheet is more stable and further away from the melting threshold than the Greenland ice sheet.

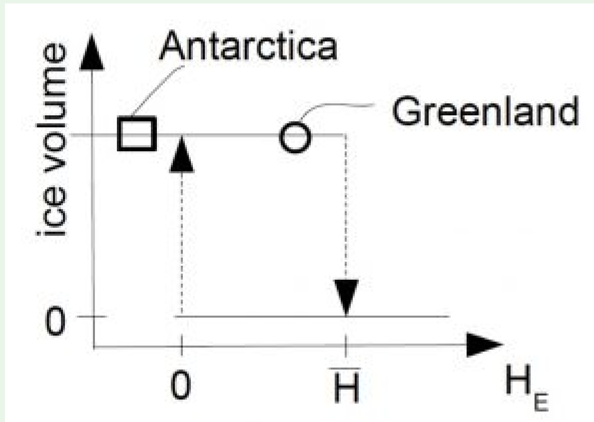


Figure B1.2: Schematic hysteresis curve for an ice sheet. The vertical and horizontal axes represent ice volume and height of the equilibrium line, respectively.

More complex ice sheet models exhibit similar threshold behavior. Current estimates suggest that the **Greenland ice sheet** may disappear almost completely for global warming of 0.8-3.2°C above pre-industrial temperatures (Fig. B1.3). However, even though this threshold may be exceeded in the near future due to continued anthropogenic carbon emissions, the ice sheet will not disintegrate immediately. It may take hundreds to thousands of years to melt completely. The larger the warming the more rapid the ice melt will be. Melting of the entire Greenland ice sheet would raise global sea level by about 7 m.

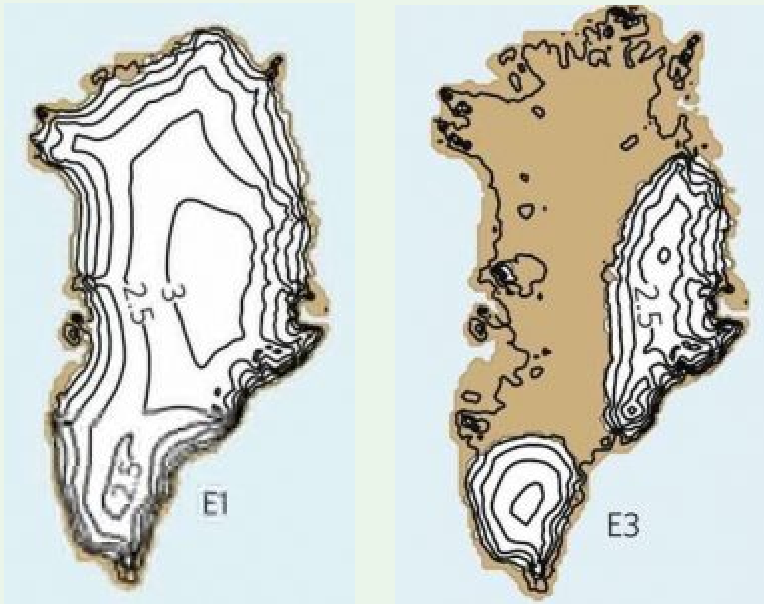
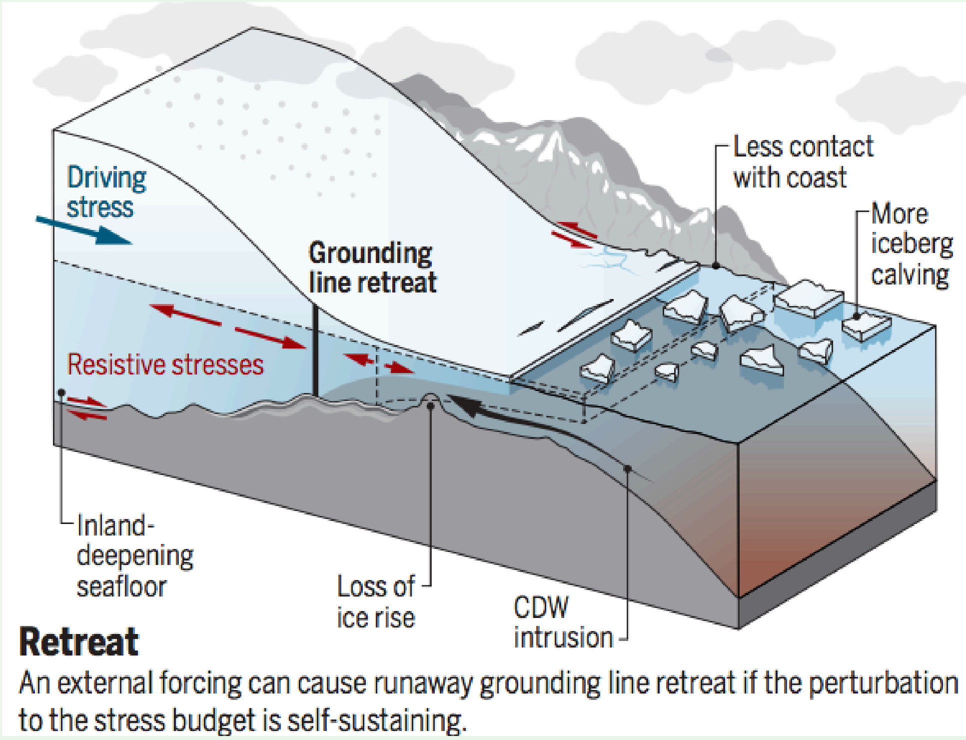


Figure B1.3: Greenland fully glaciated (left) and mostly deglaciated (right). Contour lines denote ice sheet thickness in km. From Robinson et al. (2012) at nature.com.

The **West Antarctic ice sheet** is also vulnerable to warming. Most of it is below sea level and it is fringed with ice shelves that float on the ocean (Fig. B1.4). The grounding line separates the ice shelf from the grounded inland ice. Ice shelves often provide a buttressing force that keeps the interior ice sheet from flowing faster into the ocean. Warm Circumpolar Deep Water (CDW) intrusion can melt the bottom of the ice shelf, increase calving of ice bergs and lead to retreat of the grounding line. Because the deepening of the bed towards the interior retreat of the grounding line and reduced buttressing can cause an increase in the outward ice flow and initiate an instability that leads to the collapse of the entire ice sheet. Recent models indicate that the West Antarctic ice sheet could disintegrate within hundreds of years for high and intermediate carbon emission scenarios, whereas it is stable for low emission scenarios (DeConto and Pollard, 2016). If the West Antarctic ice sheet collapsed global sea level would rise by about 5 m. Here is a short video about this topic.



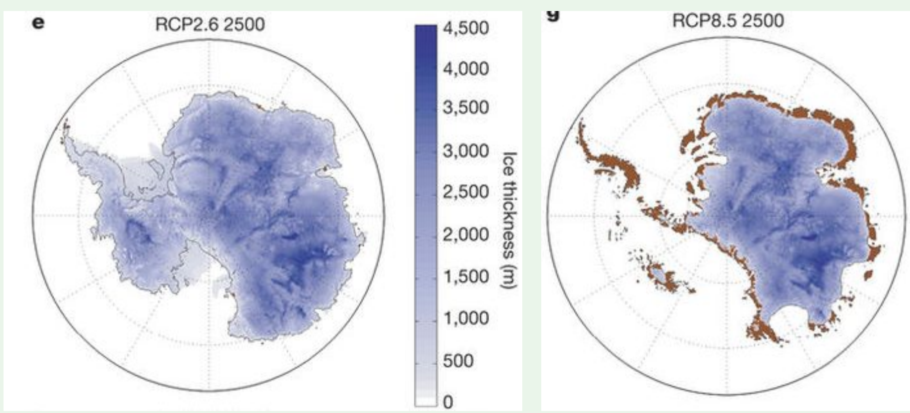


Figure B1.4: Antarctic ice sheet vulnerabilities. Top: Schematic illustration of the West Antarctic ice sheet from carbonbrief.org. Bottom: Projection of the Antarctic ice sheet at year 2500 for emission scenarios RCP2.6 (left) and RCP8.5 (right) from DeConto and Pollard, 2016).

The **Atlantic Meridional Overturning Circulation** (AMOC) also has the potential to switch into a collapsed state once a threshold is exceeded. Paleoclimate evidence indicates that such shifts happened repeatedly in the past. The AMOC exhibits hysteresis behavior because of a positive feedback between the circulation and salinities in the North Atlantic. The circulation transports high saline waters from the subtropics towards the North Atlantic. However, a reduction of the flow and thus the northward transport of saline waters will lead to decreasing salinities in the North Atlantic regions of deep water formation and this will reduce the flow even further. Currently it is thought that an AMOC collapse is unlikely for low emission scenarios but the probability increases for high emission scenarios (Bakker et al., 2016). A collapse or large reduction of the AMOC would impact not only the North Atlantic due to reduced northward heat transport, it would also affect the tropics by shifting the Intertropical Convergence Zone towards the south, which would affect precipitation patterns, vegetation, ecosystems and humans there and it would affect ocean biogeochemical cycles (e.g. of nutrients and carbon) and ecosystems.

Ecosystems will continue their poleward and upward shifts. Present day tundra, for example, will be replaced by taiga (Fig. B1.5). Species that used to live at high latitudes or on the upper elevations of mountains may be replaced by species moving in from further south or from lower altitudes. This can lead to **species extinctions**. Especially vulnerable are species that live near the top of mountains such as the cute little pika, or near disappearing sea ice such as polar bears, walrus, and narwhales (Larsen et al., 2014).

Figure B1.5: Tundra (left) is a biome at far northern latitudes characterized by grasses and shrub vegetation and permafrost beneath the seasonally thawed soil. Musk oxen, geese, caribou (reindeer), Arctic fox, Arctic hare, snow owl, and lemmings are some of the animals that call the tundra home. Warming leads to taiga (a.k.a. boreal forest, right) with coniferous trees replacing the tundra.

b) Ecosystems

Vegetation models project further poleward shifts in biomes (Fig. 10) due to warming consistent with the currently observed ongoing trends. Potential vegetation is the natural vegetation that would grow in a certain area without human land use changes such as agriculture or forestry. The biome shifts are projected with the most confidence in those regions close to the biome boundaries towards the migration direction of the invading biome. E.g. close to the tundra/taiga boundary there will be almost certainly a shift from tundra to taiga. For individual species such as aspen in North America large shifts are possible.

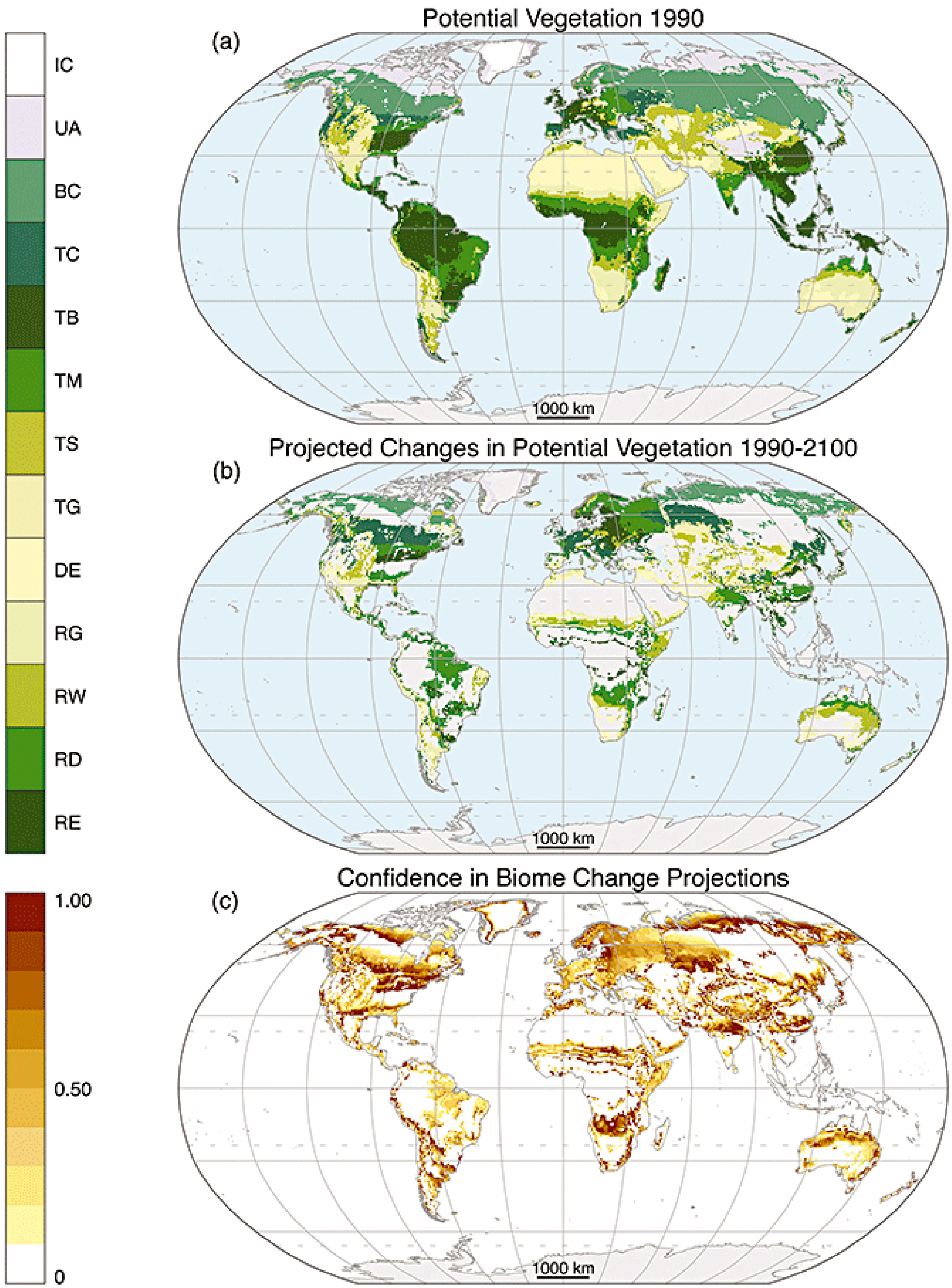


Figure 10: Modeled potential vegetation in 1990 (a), projected changes by the end of the century (b) and the confidence in the projections (c). Biomes from poles to equator are ice (IC), tundra and alpine (UA), boreal conifer forest (taiga, BC), temperate conifer forest (TC), temperate broadleaf forest (TB), temperate mixed forest (TM), temperate shrubland (TS), temperate grassland (TG), desert (DE), tropical grassland (RG), tropical woodland (RW), tropical deciduous broadleaf forest (RD), tropical evergreen broadleaf forest (RE). From Gonzalez et al. (2010).

Increasing CO_2 concentrations in the atmosphere will increase the water-use efficiency of plants. They can grow more with the same water use because they don't need to open their stomata as much for the same amount of CO_2 to flux in. They can also grow in regions that were previously too dry for them under lower CO_2. This greening effect is currently observed e.g. from satellites.

Changes in precipitation will also impact vegetation distribution. The effects on vegetation of reduced precipitation in the subtropics in a warmer climate may be compensated to some degree by the increased water-use efficiency under higher CO_2. Longer growing season, higher CO_2, and warmer temperatures will increase net plant production but also respiration, e.g. in soils.

Wildfires are difficult to project but in most regions they are likely to get worse due to climate change (Fig. 11). Fire needs fuel, which in the case of wildfire is burnable biomass (dry wood or grass). It also needs the right climatic or weather conditions. Hot and dry summers increase fire hazards and strong winds can lead to rapidly spreading fires. In many regions, e.g. North America, Australia and Russia, fire suppression by humans have caused a build up of fuel, which, once ignited, can lead to larger fires. In many parts of the tropics, on the other hand, humans burn to clear agricultural fields. Thus, changes in wildfires are not only caused by climate change. However, hotter drier summers, projected e.g. for the western United States, will lead to increased fire probability in the future. This is consistent with scientific evidence that fires have already increased there in recent decades due to anthropogenic climate change (Abatzoglou and Williams, 2016). In other regions such as the high northern latitudes, where woody taiga vegetation will replace tundra the resulting increase in available fuel may lead to more or larger fires. Note that the changes in projected fire frequency are highly uncertain as indicated by different models and methods yielding different results.

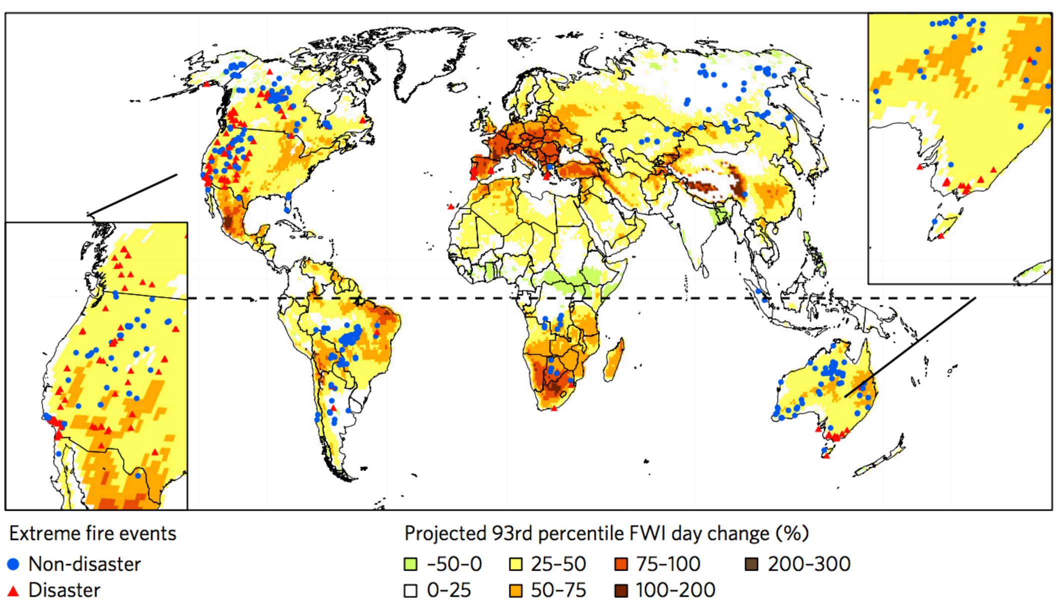


Figure 11: Projected changes in the number of days exceeding the 93rd percentile of the Fire Weather Index (FWI) by the mid 21st century (2041-2070) under a high emissions scenario (RCP8.5). Dark red shading indicates the largest increases, while the pale green shows small decreases. Red triangles and blue dots show recent extreme wildfire events. Source: Bowman et al. (2017). From carbonbrief.com.

One of the most sensitive **ocean ecosystems** are probably coral reefs. Corals are threatened not only by bleaching, which refers to the expulsion of their symbiotic algae resulting from warmer waters, but also by ocean acidification, which inhibits their calcium carbonate production. The presently observed shift of fish species towards higher latitudes will most likely continue.

Figure 11: Healthy (left) and bleached (right) coral.

In the Arctic narwhales may be replaced by invading killer whales (Fig. 12). Similarly, for many other species from phytoplankton to whales, some will benefit and others will suffer from climate change. Calcifiers such as coccolithophores will presumably be among the losers due to ocean acidification, whereas cyanobacteria, which is another group of phytoplankton, may be among the winners (Dutkiewitcz et al., 2015). Because of this complex response, many consequences for ecosystems cannot be predicted. It is therefore likely that there will be surprises that scientists could not foresee.

Figure 12: Narwhal (left) and killer whales (right).

Generally, impacts of climate change on Earth's ecosystems will be larger for high emission scenarios and smaller for low emission scenarios. Earth's ecosystems are resilient. They have experienced large climate changes in the past, such as during the ice age cycles of the Pleistocene. However, current rates of change and those expected for the future are larger for some climate variables than past rates (Fig. 13), which has many scientists concerned about the adaptability of ecosystems.

c) Long-Term Changes

Due to the long lifetime of carbon in the Earth system current human activities will affect many future generations. The ultimate fate of anthropogenic carbon is burial in ocean sediments. This is a slow process that takes tens of thousands of years to remove all of the extra carbon. For this reason much of the carbon we're putting into the atmosphere today will impact climate and Earth's physical, biological and human systems for a long time. The total amount of available fossil fuels still in the ground is uncertain (~10,000 PgC, GEA, 2012), but it is clear that enough exists to melt all major ice sheets, which would raise sea level by about 65 m. Fig. 13 shows scenarios of up to about 5,000 GtC, which leads to complete melting of the Greenland and West Antarctic ice sheets and most of the East Antarctic ice sheet with a sea level rise of about 50 m. Even the relatively low (1,280 GtC) emission scenario would lead to a long term sea level rise of about 25 m. However, the rate of change would be much slower for the lower emission scenario (0.5 m/century) compared with the higher scenario (3 m/century). Such rates are unprecedented in more than 8,000 years. High emission scenarios would melt essentially all mountain glaciers on Earth, whereas low emission scenarios would melt about 70% of current glaciers, mostly within this and the next century (Marzeion et al., 2012). Global warming for all but the low emission scenarios will be similar to or even exceed that from the Last Glacial Maximum to the early Holocene (~4°C). Transformations of Earth's ecosystems similar in magnitude to those documented for the last deglaciation can therefore be expected for intermediate and high emission scenarios. Because of the prominence of the human influence the period since the industrial revolution has been called the Anthropocene.

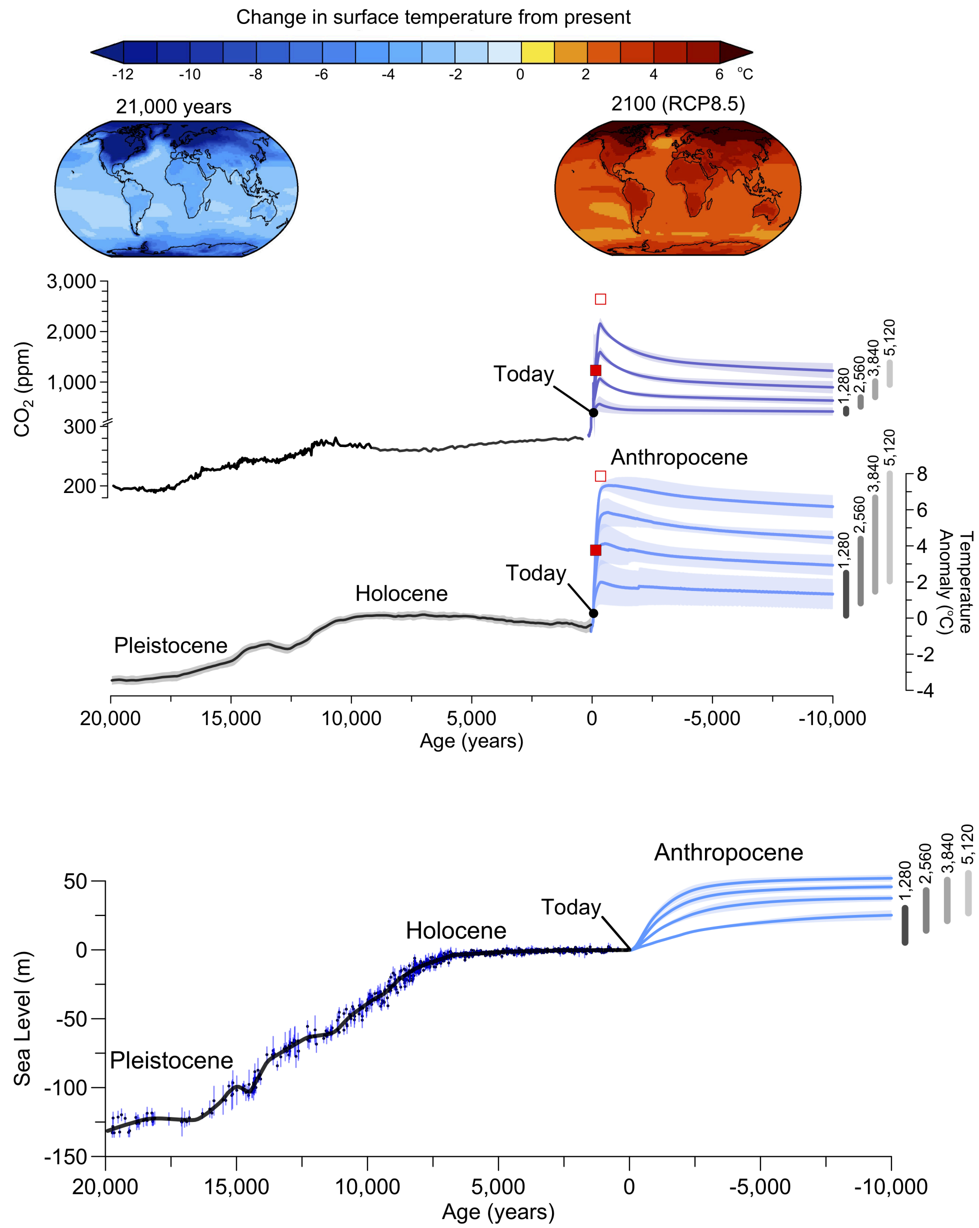


Figure 13: Long-term projected future in the perspective of past changes in atmospheric CO_2, surface temperature and sea level. Top: maps show model simulated temperature anomalies (with respect to the preindustrial) for the Last Glacial Maximum (21,000 years ago at the end of the Pleistocene) and projections for the year 2100 based on the RCP8.5 emission scenario. Center and bottom: changes in CO2, global surface temperature and sea level from paleo data and model projections. Intermediate complexity models were used for the future projections assuming a total of 1,280, 2,560, 3840, and 5,120 PgC emissions shown in the blue lines with shading indicating uncertainty. Red squares indicate results from years 2100 (solid) and 2300 (open) for the RCP8.5 scenario for comparison. Vertical gray bars show the range of uncertainty based on a range for climate sensitivity from 1.5 to 4.5°C. Modified from Clark et al. (2016). Figure courtesy of Shaun Marcott and Peter Clark.

Sea level rise will have tremendous effects on people living in coastal regions. Clark et al. (2016)

estimate that their low emission scenario (1,280 PgC) would submerge an area where currently 1.3 billion people live (19% of the global population) including 25 megacities such as Calcutta, New York, Tokyo, Shanghai, and Cairo. Due to the time lag associated with sea level rise we are already committed to future sea level rise from past emissions. By the year 2000, e.g. humans had emitted ~470 PgC and were committed to a sea level rise of about 2 m. Releasing another 470 PgC would commit us to another 9 m of long-term sea level rise.

d) Regional Changes

Projected climate impacts vary strongly depending on the region. Small islands and coastal regions will be affected by sea level rise. Polar and alpine regions will be affected by snow and ice loss. Dry regions such as the subtropics will get drier. Detailed regional projections exist for some regions such as North America and Europe. A good resource for regional impacts is part B of this IPCC (2014) report. Here we will only look at one example, which is the western United States of America. Fig. 13 shows that the snowpack there has decreased dramatically during the past 60 years. On average the April 1st snowpack has declined by 15-30% or 25-30 km^3, which is similar in volume to the region's largest man-made reservoir, Lake Mead. Further decreases in snowpack in this region can be expected due to continued warming with impacts on summer stream flows, agriculture and other human water use downstream.

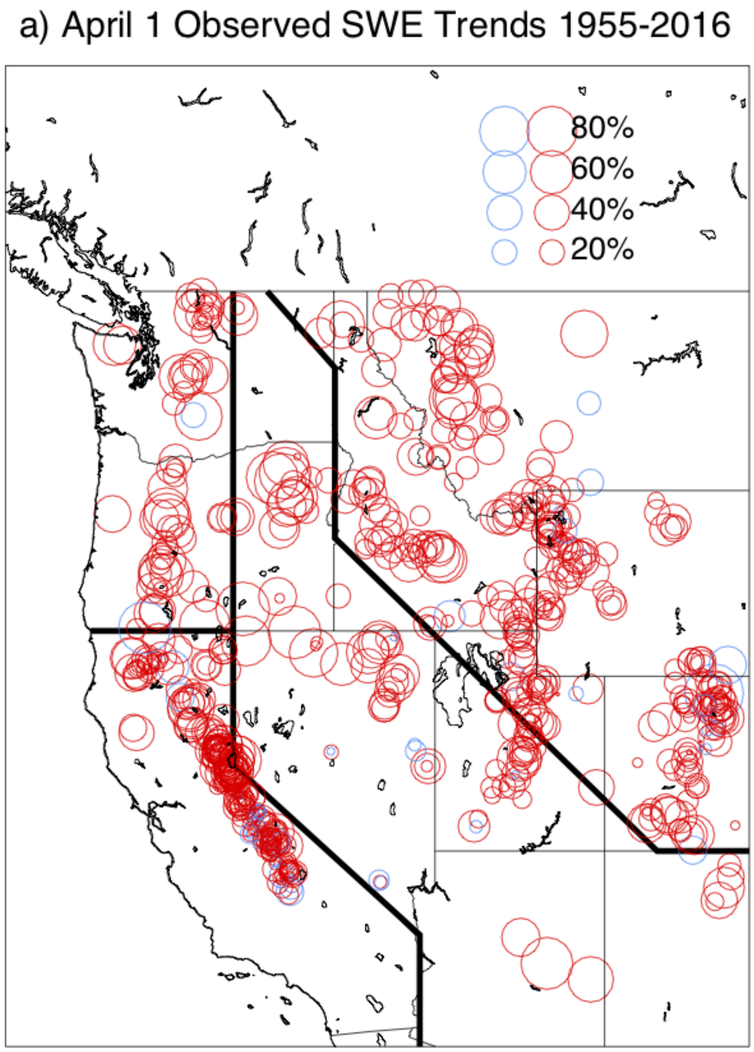


Figure 13: Linear trends in April 1st Snow Water Equivalent (SWE) observed for the period 1955-2016. Red circles indicate decreased snowpack, blue increased. From [Mote et al. (2018)](#).

Fig. (14) illustrates that vegetation west of the Cascades mountain range in Oregon and Washington is projected to shift from conifer to more drought-tolerant mixed forest, whereas some of the grassland that occupies the eastern portions of those states as well as parts of Idaho and Montana is projected to be replaced by shrubs and conifer forest. Fires are projected to increase in frequency from 37 to 9 years on average in the full region. Modeled fires for the past are infrequent west of the Cascades (~80 year recurrence time on average), but they are projected to become much more common for high emission scenarios (~27 year recurrence time for RCP8.5) due to warmer and dryer summers. Intermediate emission scenarios lead to less dramatic changes in fire and vegetation.

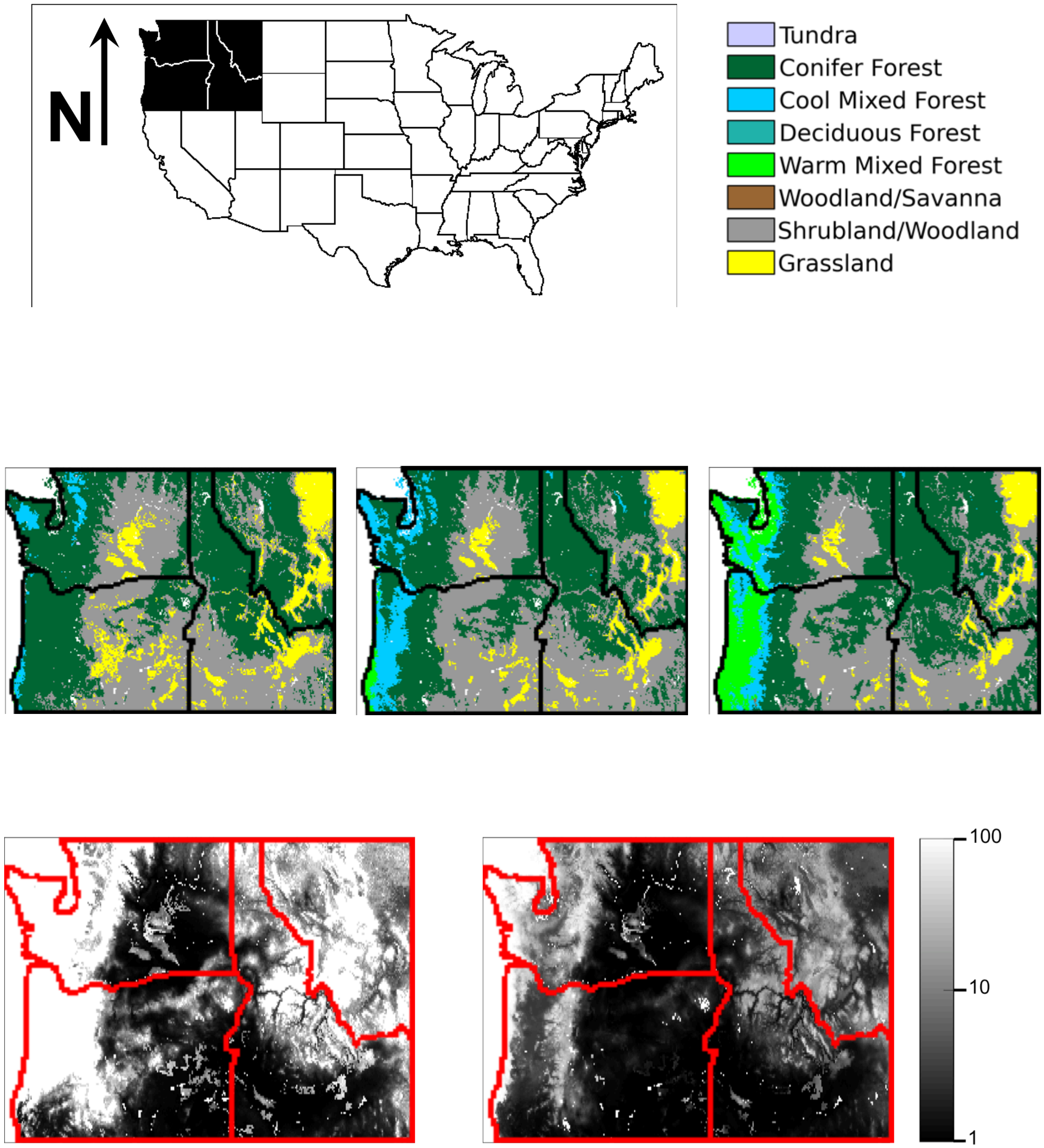


Figure 14: Projected changes in vegetation (center row with legend in top right) and fire frequency (bottom) in the Pacific Northwest (filled area in the top left map of the USA). Vegetation is shown for years 1971-2000 (left), 2036-2065 (center) and 2071-2100 (right). Fire frequency in years is shown for the 20th century (left) and the 21st century (right). Modified from Sheehan et al. (2015). Simulations without fire suppression are shown. Figure courtesy of Tim Sheehan.

e) Extremes

Figure 15: Satellite Image of Hurricane Katrina.

A warming climate implies a shift in the probability distribution such that hot extremes become more frequent and cold extremes become less frequent. This is what is currently observed and we can expect this trend to continue into the future. Due to the intensification of the hydrological cycle we can also expect more droughts and more floods. Changes in other extreme events are less well understood. Occurrences of total hurricanes and typhoons (tropical cyclones) are projected to decrease but the strongest hurricanes are projected to become more frequent (Knutson et al., 2010). This is a concern because those are the ones that inflict the most damage. Hurricane development depends on warm ocean water as an energy source. Latent heat release is also an important fuel for hurricanes and other storms. Thus warmer sea surface temperatures and more latent heat release due to more water vapor in the warmer air will strengthen hurricanes, consistent with observations of increases in the destructiveness of tropical storms in the past decades (Emanuel, 2005). Another important factor for hurricane development, particularly in its early stages, is wind shear. That is how fast winds increase with elevation. Storms require low shear to develop into a coherent vortex, which may become less likely in the future. The end result of changes in wind shear and changes in temperatures is that in a warmer climate the total number of hurricanes will decrease, but strong hurricanes will become more frequent and they will cause more rainfall.

f) Impacts on Humans

Climate change is already affecting people and more changes can be expected for the future. The decline of sea ice in the Arctic, for example, already affects native people there such as the Inuit. Their life depends on sea ice for hunting and travel and will be greatly affected by the expected future sea ice loss (Watt-Cloutier, 2015; watch her talk here). Sea ice also dampens waves and their effect on erosion of shorelines. Some villages are already threatened by increased erosion due to sea ice loss. On the other hand, a seasonally ice free Arctic will allow for ships to take a shortcut through the Northwest Passage and reduce travel time between the North Atlantic and North Pacific.

Similarly, some mining companies are already planning operations in Greenland in areas that were previously inaccessible due to ice cover. But this opportunity for resource extraction comes of course at the price of sea level rise, which will affect millions of people who live close to shore. Many people will also be affected by the loss of mountain glaciers and their summer water supply. Many current ski areas may no longer be viable in the future due to less snow cover and a shorter season, which will result in job losses there.

Climate change can also lead to conflict. The ongoing civil war in Syria, for example, has been partly caused by a drought, which has been attributed to anthropogenic climate change (Kelley et al., 2015). This attribution is in line with climate model projections that indicate dryer conditions around the Mediterranean in a warmer world (Fig. 6) and paleoclimate data (Cook et al., 2016). Violent conflict about freshwater resources is not new in the history of this and other regions but the current conflict may be the first that is partly caused by anthropogenic climate change. The resulting migrants, many of which fled to central Europe, may be one of the first climate refugees. Conflict and climate refugees may be also be expected from people displaced by sea level rise, such as in Bangladesh or low laying islands. Sea level rise affects people not only through flooding but also by saltwater seeping into the freshwater lens that exists below the ground and provides islanders often with the only source of freshwater. This can also be a problem in other coastal regions.

Questions

- What is the difference between prediction and projection?
- List two uncertainties in climate projections.
- Which tool do climate scientists use for climate projections?
- Are climate models perfect?
- Describe scenarios RCP2.6 and RCP8.5 in terms of carbon emissions and climate change. For which scenario will the impacts be greater?
- How does the global average surface temperature change projected for the end of the 22nd century compare with that between the Last Glacial Maximum and the preindustrial?
- Is projected surface temperature change larger on land or on the ocean? What are the reasons for these differences?
- Where is projected surface temperature change larger, in the tropics or at the poles? What are the reasons for these differences?
- Where are atmospheric temperatures projected to increase most in the tropics? At the surface or at high elevations? What is the reason for this?
- Where is precipitation projected to increase, where to decrease?
- How is evaporation projected to change?
- How is Arctic sea ice projected to change?
- List five different possible impacts of future climate change.
- How much carbon have humans emitted so far, cumulatively?
- How much carbon can we emit in the future if we want to stay below 2°C above preindustrial global mean temperature?

- What is a tipping point?
- Which ice sheet is closer to a tipping point, Greenland or East Antarctica?
- Why is the West Antarctic ice sheet vulnerable to a warming climate?
- How is the tundra ecosystem projected to change in the future?
- What will be some of the long-term global impacts of past and future anthropogenic carbon emissions?
- What will be some of the regional impacts in the Pacific Northwest?
- How would you expect the frequency of heat waves and cold spells change in the future?
- How are hurricanes expected to change?
- How do you expect future climate change to impacts humans?

Explore Impacts

Explore regional impacts of possible future climate change using this websites:

Videos

Lecture: Projections
Lecture: Impacts
Lecture: Tipping Points

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9. Economics

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Economics and the climate change challenge: Understanding incentives and policies

As the earth warms, the impacts are expected to affect both ecosystems and humans. Thus, understanding the impacts of climate change requires a combined understanding of how ecosystems respond to the buildup of greenhouse gases and how humans are impacted and thus respond to the changes in these ecosystems. This chapter provides an economic perspective on the climate change challenge and an introduction to the role that market-based incentives and policy can play in helping us mitigate and adapt to the impacts of climate change.

To better understand the economic levers that can be used to address our climate change problem, we can think of this challenge as similar to other decisions and outcomes that we encounter in our every-day lives and how incentives are used to alter behavior. For example, our society has reduced the incidents of lung cancer by implementing a variety of policies aimed at changing consumer behavior such as employing education campaigns that communicate the link between smoking and lung cancer, imposing taxes on cigarettes and tobacco products, banning smoking in most public places, and prohibiting the purchase by minors. While none of these are perfect deterrents for everyone, the collective outcome has been to greatly reduce the adverse impacts of smoking behaviors.

These same principles can be applied to climate change, water pollution, and other environmental challenges. In all of these applications, the environment (atmosphere, water) is viewed as an asset that provides a variety of services that support life and sustain our existence. As with all long term assets, we seek to use them sustainably. And, as with all assets, there is a value associated with their services. The value will decline as the asset is rendered less productive. Polluting these environmental assets will also decrease the level of services they can provide now and in the future. CO_2 is a form of pollution into the "atmosphere asset," where high levels of CO_2 emissions causes serious and irreversible adverse impacts. The challenges of reducing CO_2 emissions is magnified since these emissions accumulate in the atmosphere over time and also disperse throughout the global atmosphere. If society wants to slow down the rate or amount of CO_2 emissions, it needs to provide incentives that discourage such emitting behavior. Economic "tools" can be used to redirect behavior toward less CO_2 generating activities, to evaluate the most cost effective policy options and incentives to sustain this behavior, and to assess the long term costs of continued delays in collective actions to reduce greenhouse gas emissions. In this chapter we focus our economic lens on policy options to address the adverse impacts of emitting too many greenhouse gases into the atmosphere. These options include designing government-mandated or voluntary-style programs and regulations.

In order to influence behaviors and thus outcomes, one must first understand the nature of the interactions between a substantially more variable climate, and the potential damages and irreversibilites. We must also have an understanding of why the problem is occurring. Once the what and why of the problem are defined, there are a variety of policies that can be used to influence production and consumption behaviors in order to reduce or mitigate the impacts of climate change.

What is the problem? As noted in earlier chapters, human induced emissions of CO_2 and other greenhouse gases is markedly adding to the naturally occurring greenhouse gases in the atmosphere. These activities include such things as the burning of fossil fuels, manufacturing, and agricultural production. The accelerated accumulation of these greenhouse gases is causing our climate to change at rates that are unprecedented. These changes are resulting in damages to the environment and creating adverse health impacts which collectively are imposing costs on society. Damages from climate change for the USA were estimated for certain sectors of the economy in a recent study to be about 1.2% of gross domestic product per 1°C increase in global temperatures, with larger damages in southern states (see also this news article). In an effort to inform decision makers about significant potential damages in the U.S. and allow the government to set priorities and manage risks, the U.S. Government Accountability Office has recently released a report on potential economic effects of climate change. These damages/costs we will refer to as the social cost of carbon.

Why is this problem occurring? These costs to society occur because there are adverse impacts associated with human activities that are not readily apparent and often indirectly associated with the activity. In the case of energy production from fossil fuels, adverse impacts are caused by the generation of CO_2. When a producer ignores the unintended side impacts of this type of energy generation, there is an implicit cost imposed upon others. Without information on the full (private plus social) cost of carbon, as reflected in the extent of the adverse impacts, and without policies that address these costs, it is too convenient to simply ignore the impacts on our ecosystems. This Ted Talk by Pavan Sukhdev "Put a value on nature!" provides an excellent and quick overview of the problem.

Let's use electricity production from coal power plants as an example of this type of "cost to society". According to the Energy Information Administration, coal power plants in the United States emitted nearly two million metric tons of carbon dioxide in 2016. These CO_2 emissions, which contribute to global warming, are the source of the negative externality associated with electricity generation from fossil fuels.

Assume you are an electricity producer and you produce electricity by burning coal. Below is a graph of the supply and demand curves for electricity (Figure 1). The downward sloping curve can be viewed as the demand (willingness to pay) for electricity; the downward sloping demand curve can also be viewed as a marginal benefit curve because each point on the curve represents the benefit of an additional unit of electricity. The upward sloping curve represents the supply of electricity which also reflects the marginal private cost of producing each unit of electricity from this coal power plant. As the owner of a firm that is producing electricity, you would produce electricity at the quantity where the marginal benefit of electricity production, as reflected in the price consumers are willing to pay, equals your marginal private cost of producing that amount of electricity. This is the point A on the graph. At this point, the quantity of electricity produced is Qp, which is the privately optimal amount of electricity. Note that at any quantity greater than Qp the marginal benefits of that level of electricity are less than the marginal private costs of producing that level of electricity.

Now let's look at it from Society's point of view. We need electricity; it is what heats and lights our homes and powers our appliances and electronic devices. However, with each unit of electricity produced from fossil fuels there is also a certain amount of carbon dioxide emitted which is imposing unwanted costs in the form of negative impacts from climate change. This means that the total cost of producing each unit of electricity generated from fossil fuels is actually higher than just the private cost of production which basically ignores these adverse impacts associated with higher levels of carbon dioxide in the atmosphere. Thus, the true cost of producing electricity from fossil fuels is more like the red curve on this graph, which reflects a higher cost of production for each level of electricity. Economists indicate this by calculating the cost as the sum of the private costs to generate and deliver the electricity to your home PLUS the social cost of carbon.

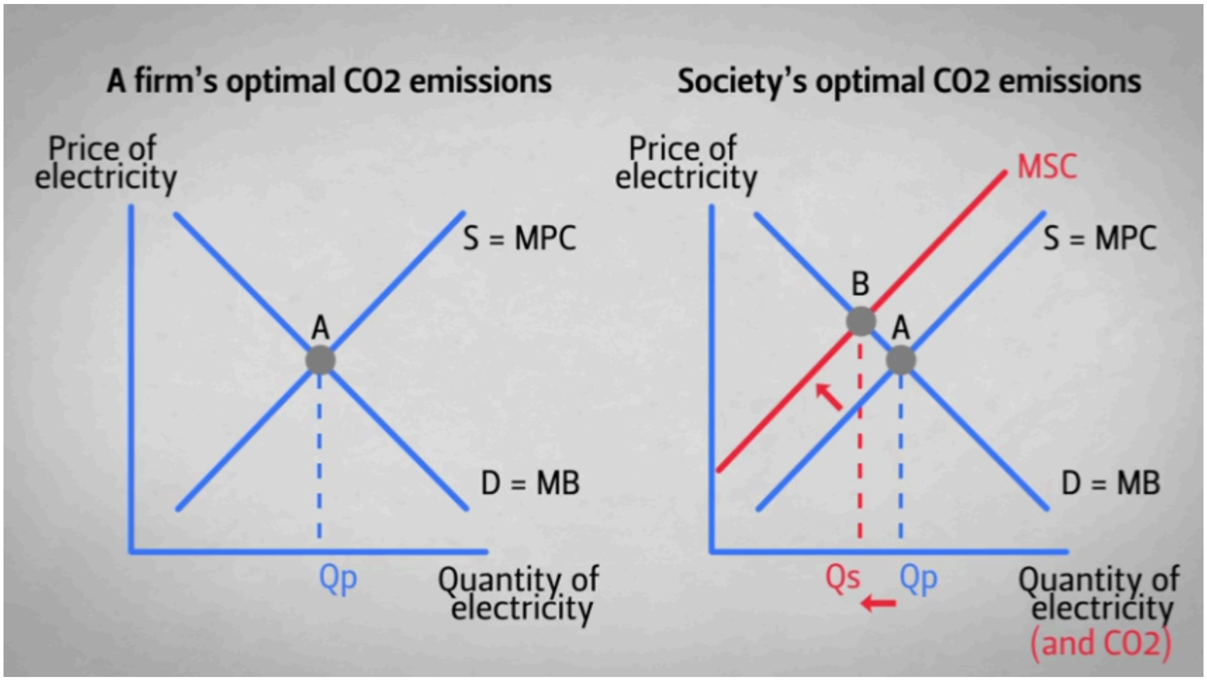


Figure 1. Optimal CO_2 emissions graphic example. See also [Climate Change Awareness Module 3: The Role of Economics](). Forward to the "Why is this problem occurring" section of the module.

Thus, in the case for electricity generated from fossil fuels (coal, natural gas), the socially optimal amount of electricity production would be Qs, which is less than the privately optimal amount Qp. In the absence of awareness of these external costs and impacts, we will continue to overproduce and overconsume electricity generated from fossil fuels. Economists refer to this as a market failure, because the (private) production and consumption of electricity exceeds the level that would be produced and consumed if the costs of CO_2 emissions were accounted for and incorporated into the total costs of production.

How can we correct this problem?

There are many options for reducing the adverse impacts to society from CO_2 and other greenhouse gas emissions, such as, switching to sources of electricity that do not contribute to the CO_2 problem, trying to generate fossil-based electricity by not releasing the CO_2 into the atmosphere, or exploring ways to reduce the demand for products that produce greenhouse gas emissions. All of these options require that the signals to producers and consumers regarding the adverse impacts are made directly, through regulations and higher prices, or indirectly through research and development on cleaner technologies. Having the ability and information to assess the benefits and the costs of the corrective actions allows decision makers to assess risks and prioritize options.

Let's see how this works using the example of the firm who is emitting CO_2 in the process of generating electricity and unaware of his adverse impacts of their actions on the accumulation of greenhouse gases. We can correct this market failure by explicitly recognizing both the private production costs and the costs to society. Economists refer to these social costs as externality costs. Since the externality costs do not show up on the electricity generating firm's expense spreadsheets they generally get ignored in the decision-making process. However, there are numerous ways to correct this cost omission. Here we present three common policies that can be used to internalize the costs of the CO_2 emissions: command and control regulations, pricing carbon through a carbon tax or a cap and trade system, and subsidies.

Command and control regulations specify how a producer must manage his/her production process that is also resulting in generation of the CO_2 pollution, establishes a monitoring procedure(s), and enforces a set of standards aimed at either the production process itself or the quantity of electricity. Let's look at our graph again (Figure 2). Here we have an illustration of the quantity of electricity that is being produced (Qp) and the socially optimal amount when carbon emissions are considered (Qs). How can we achieve the reduction from Qp to Qs?

One way would be to impose an emission standard which dictates that the quantity of electricity produced shall not to exceed Qs. If a producer generates more than this amount, a fine is imposed. When the fine is set high enough the producer will choose to reduce production rather than incur the fine, thus achieving the socially optimal quantity of electricity (Qs). Note that the emission standard is indirectly reflected in the graph below since emissions are tied to electricity production. This also assumes that we have good information and can determine the exact damages from CO_2 emissions, and we can accurately measure CO2 emissions associated with the production of electricity. It is a tall order, but with current technologies and monitoring process it is not impossible.

Determining the precise emissions standard, however, is difficult. If the emission standard is too restrictive and results in production levels to the left of Qs, the policy is overly costly. On the other hand, if the emission standard is too lax, the target for emissions reductions that are necessary will not be met.

In the United States, command and control policies are often used by the Environmental Protection Agency to ensure clean air and water. For example, under the Clean Water Act, there are many waterways that have Total Maximum Daily Loads standards that set a limit on the amount of pollutants that can occur in water ways. Likewise, the Clean Air Act created National Ambient Air Quality Standards (NAAQS) for many pollutants that are harmful to both public health and the environment such as Sulfur Dioxide, lead, nitrogen dioxide and even particulate matter, which can cause breathing problems. These command and control policies have been highly successful in improving the water and air quality in the United States. However, some argue that there are other more efficient ways to obtain the same outcome, and that command-and-control regulations do not provide incentives to improve beyond the standard set (Tietenberg 1985, Stewart 1996).

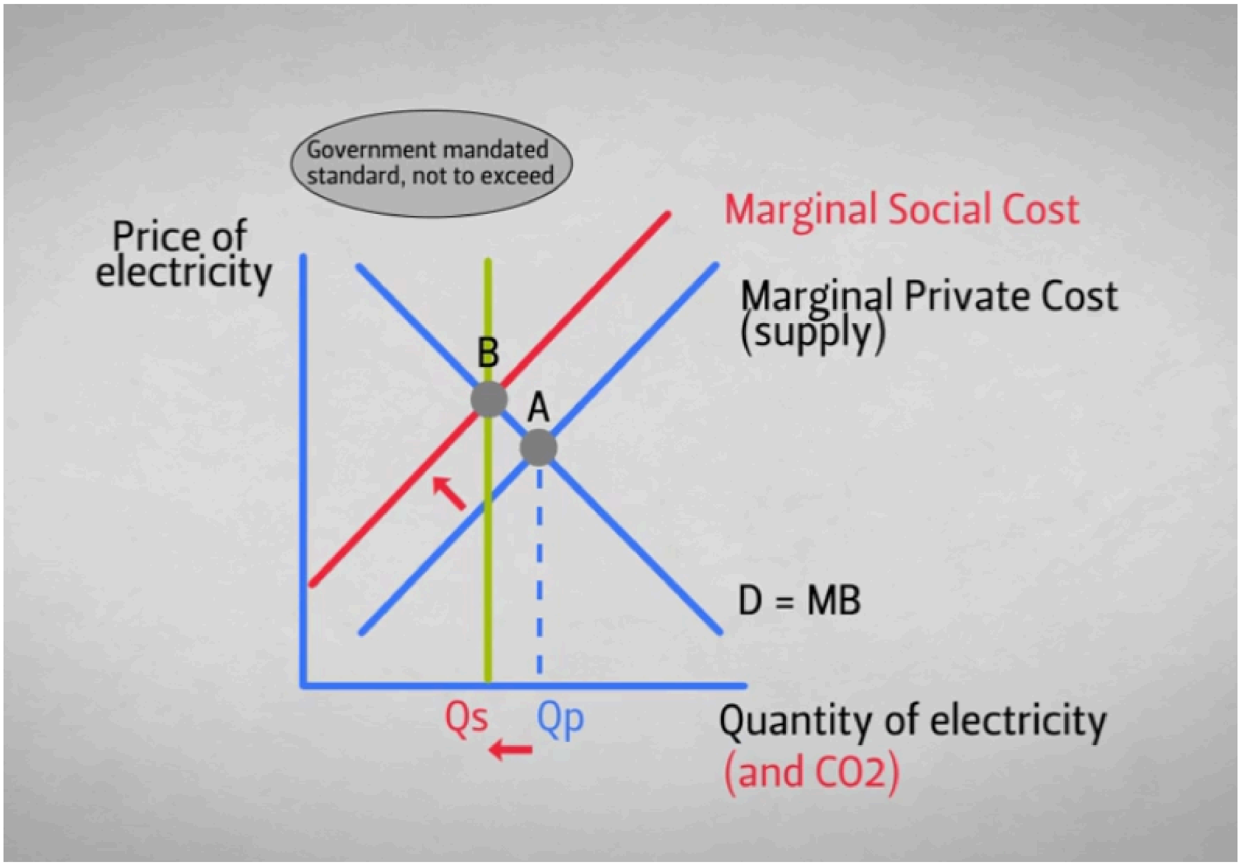


Figure 2. Command and control graphic example. See also [Climate Change Awareness Module 3: The Role of Economics.]() Forward to the "How can we correct this problem" section of the module and click on the Command-and-Control regulation button for a more detailed explanation of command-and-control regulations.

Putting a **price on carbon** is the method many economists favor for reducing or controlling greenhouse gas emissions. Pricing carbon provides producers and consumers with a monetary incentive to reduce greenhouse gas emissions by placing a value on each unit of carbon dioxide or carbon dioxide equivalent that is emitted into the atmosphere. The carbon price can be viewed as the amount that must be paid for the right or permission to emit one unit of carbon dioxide into the atmosphere. This is a direct way to incorporate the costs to society of greenhouse gas emissions into the decision-making process of producers and consumers. Carbon pricing is usually either in the form of a tax or a combination of a cap on emissions with the ability to trade carbon dioxide emission allowances, referred to as a cap-and-trade system.

A **carbon tax** is an incentive that encourages companies and households to invest in cleaner technologies and adopt greener practices, by increasing the price of an item that contributes to the buildup of greenhouse gases into the atmosphere. If the tax on the greenhouse gas emissions is set high enough, the increased price for the product that is produced using a technology that generates greenhouse gases as a by-product of electricity, provides an incentive for producers to reduce emissions, by either reducing production, or investing in technologies that produce less carbon or that capture the carbon. Often a proportion of this increased cost is passed to consumers in the form of higher prices, which also incentivizes consumers to purchase less.

In Figure 3 below we illustrate how a tax can be used to incorporate the cost to society of electricity production, and reduce electricity production and the associated CO_2 emissions from Qp to Qs. Recall from our earlier discussion, in a situation where there is no standard and no carbon tax, the quantity of electricity that firms produce will be Qp. When we add a carbon tax to the price of each unit of electricity, we shift the supply curve to the left. This is represented by the marginal social cost curve represented by the red line in this graph. At the new equilibrium where marginal social cost is equal to marginal benefits, the quantity supplied shifts from Qp to Qs just as in the Command and Control case. However, in this case a tax in the amount represented by the tan box is collected by the federal government. The government can then use this tax in a variety of ways such as giving the tax revenues back to the general population as a tax refund, investing in research on or construction of carbon reduction technologies, or using the money to reduce budget shortfalls.

Carbon taxes have been implemented in many countries, such as Finland, Denmark, Norway, Sweden, and British Columbia, among others. Success is often measured as reduced emissions, but other factors such as technology innovation and industrial efficiency gains have also been cited as factors of success. To learn more about carbon taxes and where they have been implemented refer to Carbontax.org. This website explains the basics of a carbon tax and provides examples of where carbon taxes have been implemented around the world.

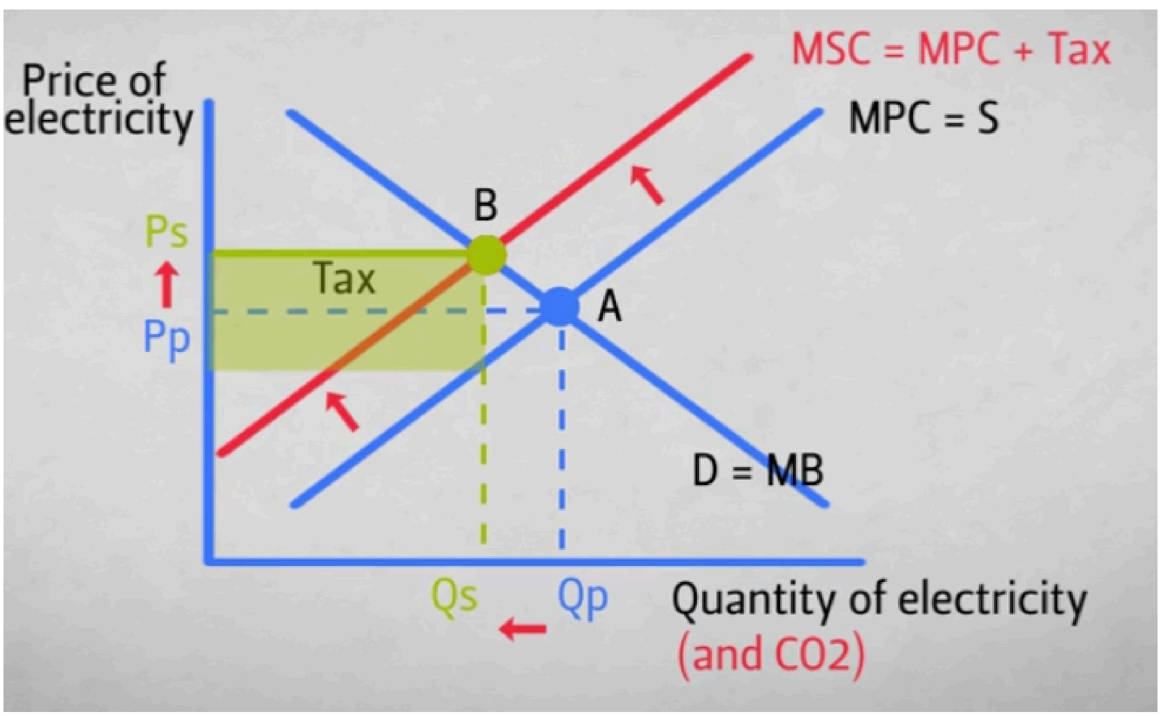


Figure 3. Carbon tax graphic example See also Climate Change Awareness Module 3: The Role of Economics. Forward to the "How can we correct this problem" section of the module and click on the Carbon Tax button for a more detailed explanation of carbon taxes.

A **cap-and-trade system** combines the command-and-control policy of setting a cap on emissions with a carbon pricing policy. With a Cap-and-Trade program, a limit or cap on emissions is set, and polluters receive or purchase emissions allowances. The total allowances are limited by the cap. Each pollution source (firm) can then design its own compliance strategy. They may choose to install pollution controls and implement efficiency measures; they also have the option to sell any excess allowances or purchase allowances if they find they cannot meet the emission standard by other means. This policy is most impactful when the firms within the industry are not identical and have different costs to reduce pollution. The key difference between this and the Command and Control policy is that the cost to each firm of compliance may be less than a taxing scheme because low-cost firms are allowed to trade allowances. Under a market based trading system, firms that can abate at a lower cost will choose to sell some of their allowances to firms that have higher costs of abatement.

The Environmental Protection Agency has successfully used cap-and-trade systems in the past to set up several clean air markets. Perhaps the most well-known is the Acid Rain program. This program has substantially reduced SO_2 and NO_x released into the atmosphere. These are compounds that can readily mix with water and oxygen in the atmosphere to form sulfuric and nitric acids, which then fall to the ground as acid rain. Acid rain is detrimental to plants, wildlife, fish, buildings, and humans. More recently cap-and-trade systems have been used to reduce carbon emissions in the Northeastern United States (Regional Greenhouse Gas Initiative) and California. For answers to some common questions about emissions trading, see these short videos about the Emissions Trading Scheme in New Zealand by Motu Research.

To better understand how a cap-and-trade program works and how it could be more efficient than command-and-control policies, let's look at an example (Figure 4). Suppose we have two plants that are each currently emitting 100 tons of carbon dioxide for a total of 200 tons. The government wants to cap emissions at 100 tons. This requires a 100 ton reduction in emissions, so they establish a command-and-control policy requiring each plant to reduce emissions by 50 tons. Plant A can reduce emissions at a cost of $10 per ton, but plant B is less efficient and it costs them $20 per ton to reduce emissions. When each firm has to reduce emissions by 50 tons, the total cost of emission reductions is $1,500.

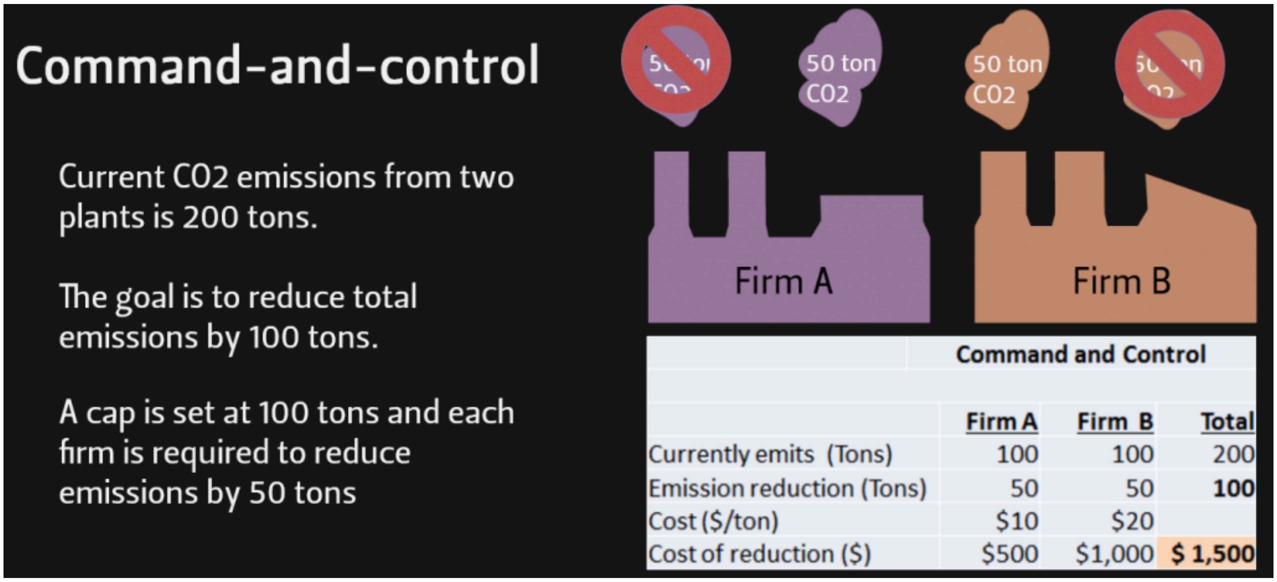


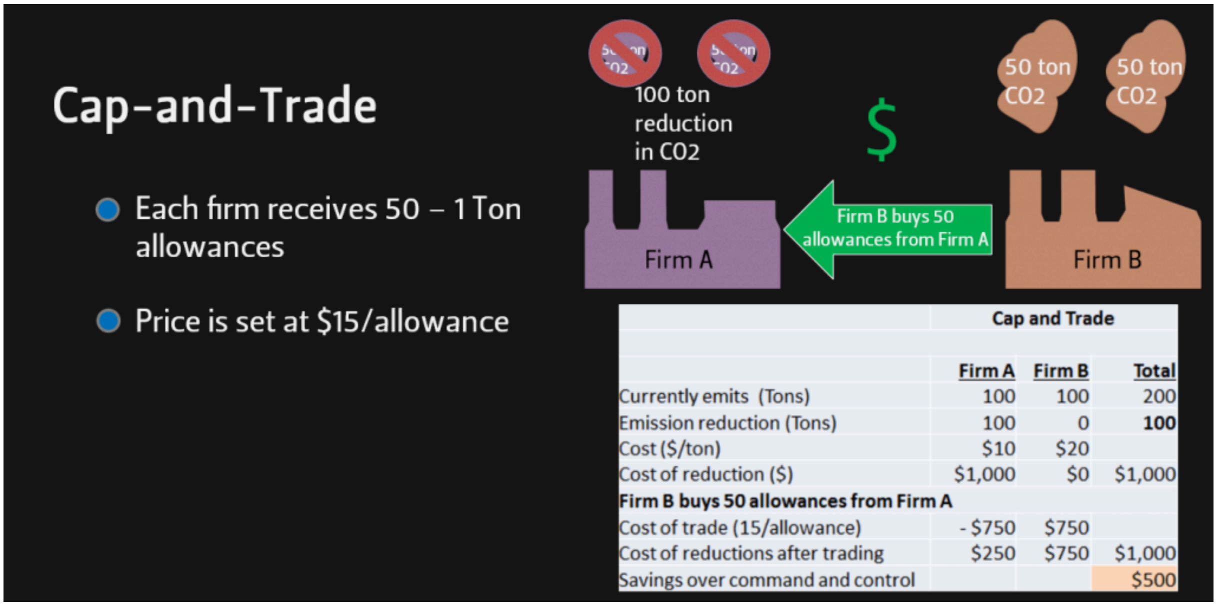
Figure 4. Command-and-control example See also Climate Change Awareness Module 3: The Role of Economics. Forward to the "Comparing command-and-control to cap-and-trade" section of the module for an illustrated comparison.

Now let's look at how the costs would differ under a cap-and-trade policy (Figure 5). In this example each firm is given 50 allowances of 1 ton each. Since it is less expensive for Plant A to reduce emissions, they would benefit from reducing emissions and selling some or all of their allowances to Plant B.

Let's assume that the price of allowances is set at $15. Just as in the previous example it costs Firm A $10/ton to reduce emissions and it costs Firm B $20/ton. Because it only costs plant A $10/ton to reduce their emissions, they will decide to reduce their emissions by 100 tons (bringing their emissions to 0). They can then sell all their allowances and receive $750. When this is subtracted from their cost of emissions reductions of $1000, their cost of emissions reductions after trading is only $250.

At $15/ton Firm B would choose not to reduce emissions at all , but to purchase 50 allowances from A for $750 because it will cost them $5/ton less than reducing their own emissions. Thus instead of paying $1000 to reduce their emissions by 50 tons, they pay $750 to purchase 50 allowances.

Even though Firm B does not reduce their emissions, the goal of 100 tons of emissions reductions is still met, because Firm A chose to reduce all of their emissions. In this case the goal of emission reductions is met with a total cost of only $1,000 a savings of $500 from the command and control option.



	Cap and Trade		
	Firm A	Firm B	Total
Currently emits (Tons)	100	100	200
Emission reduction (Tons)	100	0	**100**
Cost ($/ton)	$10	$20	
Cost of reduction ($)	$1,000	$0	$1,000
Firm B buys 50 allowances from Firm A			
Cost of trade (15/allowance)	- $750	$750	
Cost of reductions after trading	$250	$750	$1,000
Savings over command and control			$500

Figure 5. Cap-and-Trade example. See also [Climate Change Awareness Module 3: The Role of Economics.]() Forward to the "Comparing command-and-control to cap-and-trade" section of the module for an illustrated comparison.

Plant A benefits financially from being more efficient and plant B has an incentive to find cheaper ways to reduce emissions in the future, so that they could benefit from selling their allowances or at least not have to purchase as many in the future.

Another common method that has been used to correct this problem is the **use of subsidies**. Subsidies are used to encourage desired behaviors. Subsidies represent the carrot approach as opposed to the stick approach of command and control policies. Subsidies come in various forms such as cash grants, low interest or interest-free loans, tax breaks, and rebate, offered to consumers and producers. For example in order to get producers and consumers to invest in energy efficient products and energy systems, Oregon has offered renewable energy development grants to organizations that plan to install renewable energy systems, as well as personal income tax credits to homeowners and renters for purchasing energy-efficient products and energy systems for their homes. Subsidies are often offered for only a short period of time to initiate changes in behavior with the aim that these changes will encourage and enhance emerging markets until they can work without government interference.

Discussion

There is no one climate policy that is the perfect solution. To make significant progress in combating climate change, a combination of several policies that influence both producer and consumer behaviors will be required. For example, as we have mentioned above, Oregon has offered subsidies in the form of tax rebates and grants. The Oregon legislature has also been looking into several carbon pricing bills which include cap and trade or cap and invest systems as well as a carbon tax and shift program. In 1996, Oregon also created a CO_2 emission standard. This required new energy facilities to meet the CO_2 standards or pay a carbon tax per metric ton of excess CO_2. Facilities also have the option to provide cogeneration that offsets fossil fuels or invest in projects that offset CO_2 emissions. Simultaneously the (Oregon) Climate Trust was established as a nonprofit organization and given the authority to purchase and retire CO_2 offsets with the taxes collected from excess CO_2 emissions.

Attempts to gain support for a national carbon tax or carbon cap-and-trade system have not yet been successful, but regional cap-and-trade systems such as the Regional Greenhouse Gas Initiative (RGGI) and the California Carbon Tax program have been showing signs of successfully controlling carbon emissions. As an added benefit, revenues from these programs have been used to offset rate increases and support investments in carbon saving technologies. For example the RGGI's 2015 report, states that 64% of RGGI 2015 investments were used to support energy efficiency programs in the region, 16% were used to fund clean and renewable energy programs, and 4% have funded GHG abatement programs. These programs have also spurred local economic growth and job creation.

As knowledge of the impacts of climate change grows, consumer demand for alternative sources of energy is also growing and creating new markets and new jobs. For instance, several companies have emerged in recent years that build and or install windmill and solar panels. In addition to creating clean energy sources, these companies also create new employment opportunities and create demand for many raw materials that are needed to build these systems. The increase demand for these products and growth of alternative energy facilities has significantly contributed to the electricity capacity in the US in recent years. For example, about 60% of electricity generation capacity added to the U.S. grid in 2016 came from wind and solar (Cusick 2017). As these new energy markets become more stable, state and local governments will be eliminating many of the subsidy programs that have helped to establish these markets. The sun setting of many Oregon Department of Energy tax credits at the end of 2017 is a prime example.

Questions

- Can you identify some of the causes of climate change? Do you think global warming is having any impact on your daily life?
- To date there remains some uncertainty regarding the impacts that policies to slow down the rate of GHG emissions may have on economic development especially in lower income countries. Why is better information on these impacts critical to the design of policies and programs to address GHG emissions? How would you respond to those who suggest that we need to postpone taking any action to address GHG emissions until we "know all the facts"?
- How does an economist characterize the difference between private and social costs of production? In thinking about the air and water as environmental resources that humans value, how would you define the private and social costs associated with a manufacturing firm that is dumping waste products into the

river, as was typical for textile firms in the 1900s in New England? Can you suggest a more recent example where the social costs of production exceed the private costs? What regulations have local and state governments imposed to address these situations?

- Are there examples where you are negatively impacted by others' behavior? In these examples, what actions could you take to reduce the adverse impacts you are experiencing? Would these be more of a command-and-control (or ask-and–hope) type of request, or a request that compensates you for your "suffering"?
- Suppose you are a legislator voting on a series of options to address greenhouse gas emissions. What would be some of the things to consider when choosing one policy over another?
- How would you change or modify your driving and transportation choices if (1) the price of gasoline increased by 30 percent? (2) the price of gasoline doubled? (3) you were offered a $2000 rebate towards an electric car? (a mileage tax was imposed on your personal vehicle each year? (4) public transportation on a bus or train was free? Do you think incentives of this sort will work in the long-run? Why or why not?

Videos

Lecture: Economics
Lecture by William Nordhaus

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Motu Research. How does the Emissions Trading Scheme Help New Zeland Reduce Climate Change? All Questions. Published on Oct 3, 2017. https://www.youtube.com/watch?v=1GtMJEQPCHk. Individual questions can also be viewed at these links:

1. What is emissions trading?
2. How does emissions trading affect me?
3. We've had an ETS for years why are emissions still going up?
4. Can the ETS be fixed?
5. What can we do about dodgy carbon credits?
6. Is the ETS enough?
7. Wouldn't a carbon tax be better?

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10. Ethics

by Kathleen Dean Moore[1] and Michael Paul Nelson[2]

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Do We Have a Moral Responsibility to Stop Global Temperature Rise?[1]

The Relevance of Ethics to the Climate Change Discussion

Scientists continue to provide overwhelming evidence that greenhouse gas pollution, environmental degradation, and consequent global climate change are profoundly dangerous to humans and to other life on Earth. A group of 500 scientists led by a team from Stanford issued this recent warning: "Unless all nations take immediate action, by the time today's children are middle-aged, the life-support systems of the Earth will be irretrievably damaged" (Barnosky et al., 2013). But to the surprise and frustration of the scientists, all nations are *not* taking immediate action to slow climate change, and people are largely silent, even acquiescent, in the face of real threats to their futures and the futures of all beings that evolved under this – not another, hotter, more volatile and violent – climate.

What accounts for this disconnect between facts and actions?

An answer can be found in the logic of practical decision-making, the form of reasoning that leads from facts to sound conclusions about what course of action a person or government should take. The logic goes like this:

Any argument that reaches a conclusion about what ought to be done will have two premises. The first is a *statement of fact*, a descriptive statement based on empirical evidence, often grounded in observation and science: *This is the way the world is; this is the way the world may become under a certain set of conditions*. The second premise is a *statement of value*, a prescriptive

1. Adapted from Moore and Nelson (2010, 2012)

statement, a moral affirmation based on cultural values and ethical norms: *This is the way the world ought to be; this is good, this is just, this is a worthy goal.* From this partnership of facts and values, but from neither alone, we can reason to a reliable conclusion about what we ought to do (Box 1). One might say that the first premise alone is a world without a compass. The second premise alone is a compass without a world. Only together can they point in the direction we ought to go.

Box 1. The Logic of Practical Decision-making. An Example

The factual premise	*If we do not act soon, anthropogenic environmental changes will bring serious harms to the future.*
The ethical premise	*We have a moral obligation to avert harms to the future, so as to leave a world as rich in life and possibility as the world we inherited.*
The conclusion	*Therefore, we have a moral obligation to act, and act now.*

Scientists have done an impressive, sometimes even heroic, job of documenting the factual premise. But the ethical premise is still under discussion – what is our obligation to the future? The stakes of this discussion are high.

Do We Have a Moral Obligation to Take Action to Protect the Future of a Planet in Peril?

A 2010 project asked one hundred of the world's moral leaders from a wide variety of worldviews and continents if we – governments and individuals – have a moral obligation to do what we can to prevent catastrophic climate change, and if so, why (Moore and Nelson, 2010). The goal was not to find the right answer, but to find a great abundance of answers, so that no matter what views people bring to the discussion, they will find at least one reason that speaks powerfully to them. Yes, the moral leaders wrote back, we must take action, for a wide variety of reasons. Here are just seven of their answers (Box 2).

Box 2. Do we have an obligation to take action to prevent catastrophic climate change?

1. *Yes, to protect the thriving of humankind.*
2. *Yes, for the sake of the children.*
3. *Yes, for the sake of the Earth and all its lives.*
4. *Yes, because the gifts of the Earth are freely given, and we are called to gratitude and*

reciprocity.

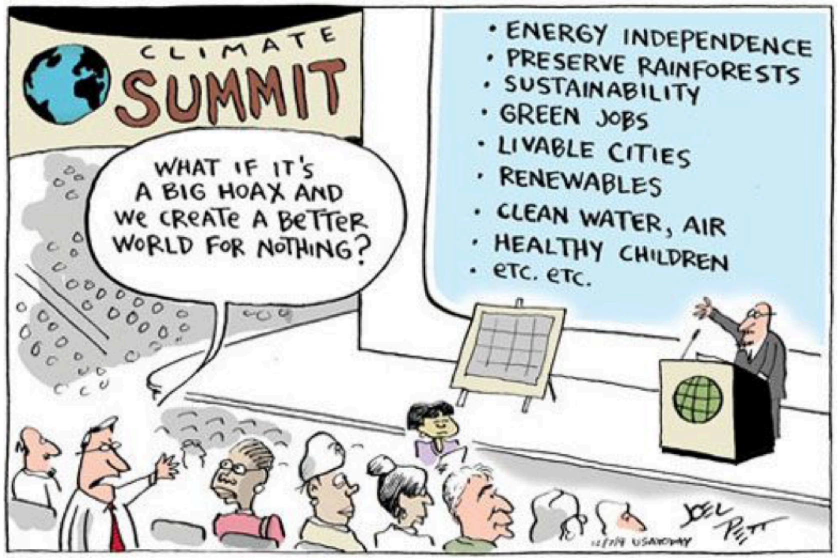
5. *Yes, because compassion requires us to reduce or prevent suffering.*
6. *Yes, because justice demands it.*
7. *Yes, because our moral integrity requires us to do what we believe is right.*

1. We must act, to protect the thriving of humankind.

Daniel Quinn, author of *Ishmael*, explained our peril. "We are like people living in the penthouse of a hundred story building. Every day we go downstairs and at random knock out 150 bricks to take upstairs to increase the size of our penthouse. Since the building below consists of millions of bricks, this seems harmless enough . . . for a single day. But for 30,000 days? Eventually—inevitably—the streams of vacancy we have created in the fabric of the walls below us must come together to produce a complete structural collapse. When this happens—if it is *allowed* to happen—we will join the general collapse, and our lofty position at the top of the structure will not save us." (Quinn, 2010)

Of course, not everyone thinks that a catastrophic crash in human numbers would be a bad thing — wouldn't the world be better off without us? But consider: for whatever purpose and by whatever process, in humans, the universe has evolved the capacity to turn and contemplate itself – to seek to understand the universe and to celebrate the mysteries of what we cannot understand. And whatever the faults of our species – and they are innumerable and tragic – we, maybe alone, have the capacity to imagine how we might be better.

This is the positive side of action to avert climate catastrophe. At this hinge point in history, we have not only the chance to escape the worst of the harms, but the chance to make a "great turning" (Macy and Johnstone, 2012) toward a healthier, more just and joyous planetary civilization.



Cartoon from Joel Pett at Cartoon Arts International.

The upshot: If severe planetary change threatens to undermine the foundations of human thriving, and if human thriving is a fundamental value, then we have an obligation to avert the degradations that threaten us. Anyone who accepts the scientific evidence about the dangers of climate change and affirms the value of human life, will not be able to sit on their hands.

2. We must act, for the sake of the children.

James Speth, former Dean of the School of Forestry and Environmental Studies at Yale, writes, "All we have to do to destroy the planet's climate and ecosystems and leave a ruined world to our children and grandchildren is to keep doing exactly what we are doing today" (Speth, 2010). If climate destabilization will be manifestly harmful to children, as Speth claims, and if we have a moral obligation to protect children, then we have an obligation to expend extraordinary effort to prevent catastrophic climate change.

Then twelve -year old Severn Suzuki, speaking at the UN's Earth Summit in Rio de Janeiro, said, "Parents should be able to comfort their children by saying 'everything's going to be all right,' 'it's not the end of the world,' and 'we're doing the best we can.' But I don't think you can say that to us anymore." The question then is: What must we do, in order to tell our children honestly that we're doing the best we can for them?

It's important to think carefully about what those extraordinary efforts are. People might say, "I don't care about ethics. All I care about are my children. And I am going to make as much money as I can, so that they can be safe and happy all their lives." Doesn't everyone want a safe and happy future for their children? The irony, of course, is that we harm them even as (especially as) we try to provide for them. In the end, the amassing of material wealth in the name of our privileged children's future is what will hurt them the most, as it exhausts the resilience of the planet's life-supporting systems. And what our decisions will do to the children who are not privileged is not just an irony; it's a moral wrong. These children, who will never know even the short-term benefits of misusing fossil fuels, are the ones who will suffer first as rising seas flood their homes, fires scorch cropland, diseases spread north, and famines scourge lands that had been abundant.

3. We must act for the sake of the Earth and all its lives, because the community of Earth and its lives has intrinsic and infinite value.

The failure to act on behalf of the Earth and all its creatures is, of course, a great imprudence – a cosmic cutting-off-the-limb-you're-sitting-on stupidity. But it is also a moral failure. That is because the planetary community (this swirling blue sphere crammed with life) is not only *instrumentally* valuable. That is, it's not just valuable because it is supportive of human life. Rather, the Earth, like a human being, has value in and of itself. It has what philosophers call *intrinsic* value. We have responsibilities to honor and protect what is of value. So we have the responsibility to honor and protect the Earth as we find it, a rare blue jewel in the solar system.

Kim Heacox, reprinted by permission.

Philosopher Kathleen Dean Moore writes,

> *Premise 1. It's not just the sun in winter, the salmon sky that lights the snow, or blue rivers through glacial ice. It's the small things, too – the kinglet's gold crown, the lacy skeletons of decaying leaves, and the way all these relate to one another in patterns that are beautiful and wondrous. The timeless unfurling of the universe has brought the Earth to a glorious richness that awakens in the human heart a sense of joy and wonder.*
>
> *Premise 2. It is right to protect what is wondrous and wrong to destroy it. This is part of what "right" means – to enhance, rather than diminish, what is of value.*
>
> *Conclusion. This is how we ought to act in the world – with respect, with deep caring and*

fierce protectiveness, and with a full sense of our obligation to the future, that this planetary richness shall remain.[2]

4. Yes, because the gifts of the Earth are freely given, and we are called to gratitude and reciprocity.

Begin with this fact: The gifts of the Earth (what we cravenly call "natural resources" or "ecosystem services") are freely given — rain, sun, fresh air, rich soil, all the abundance that nourishes our lives and spirits. Perhaps they are given to us by God or the gods; maybe they are the fruits of a fecund Earth. It doesn't matter to the argument: let that be a mystery, why we are chosen to receive such amazing gifts. What is important is that they are given. We do not earn these gifts. We have no claim on them. If they were taken away, there is nothing we could do to retrieve them. At the same time, we are utterly dependent on these gifts. Without them, we quickly die. This unequal relationship, the relationship of giver and receiver of gifts, makes all the moral difference.

We understand the ethics of gift-giving. To receive a gift requires us to be grateful. To dishonor or disregard the gift — to ruin it, or waste it, to turn it against the giver or lay greedy claim to it or sourly complain — all these violate our responsibilities as a recipient. Rather, to be grateful is to honor the gift in our words and our actions, to say, "This is a great gift," and to protect it and use it well. In this way, gratitude calls us to attentiveness, celebration, and careful use.

Furthermore, an important part of gratitude is reciprocity, the responsibility to give in return. We give in return when we use our gifts well for the benefit of the Earth and the inhabitants who depend on its generosity. In this way, gratitude for our abundant gifts is the root of our moral obligation to the future to avert the coming climate calamities and leave a world as rich in possibilities as the world that has been given to us.

5. We must act from compassion, which requires us to reduce or prevent suffering.

Of all the virtues that a human being can possess, the greatest may be compassion. 'Compassion,' to 'feel with,' to imagine ourselves in another's place. To be frightened as they are frightened by a suddenly unstable world, to be bewildered as they wonder where to turn, to suffer their thirst and anger. Understanding the joys or sufferings of others, the compassionate person is joyous or

2. Moore and Nelson (2010) 329-330.

suffers too. The truly compassionate person also acts in the world, providing conditions that bring forth joy and preventing or diminishing conditions that create pain.

Among the calamities of climate change and the resulting environmental degradation is an increase in human suffering and the suffering of other feeling beings. Climate change disrupts food supplies, reduces or contaminates drinking water, spreads disease, increases the terror of storms, floods great cities, and cracks villages into the sea. The price of the reckless use of fossil fuels will be paid in large part by human suffering.

Unknown artist.

If virtuous people are compassionate, if compassionate people act to reduce suffering, if climate change will cause suffering greater than the world has ever known, then we who call ourselves virtuous have an inescapable obligation to the future to avert the effects of the coming calamities.

6. We must act, because justice demands it.

If people have inalienable rights to life, liberty, and the pursuit of happiness, then the carbon-spewing nations are embarking on the greatest violation of human rights (Universal Declaration of Human Rights) the world has ever seen. Uprooting people from their homes, exposing them to new disease vectors, disrupting food supply chains — it's a systematic violation of human rights. By whom, and for what? By the wealthy nations who can't or won't stop spewing carbon into the air. For what? For self-enrichment, the continuation of wasteful and pointless consumption of material goods. Why? Because of the failure of conscience or will to create a fairer way of living on the planet.

It's not just a violation of rights: Those who are suffering, and will suffer, the most severe harms from climate change (at least in the short term, until it engulfs us all) are those least responsible for causing the harm. That's not fair.

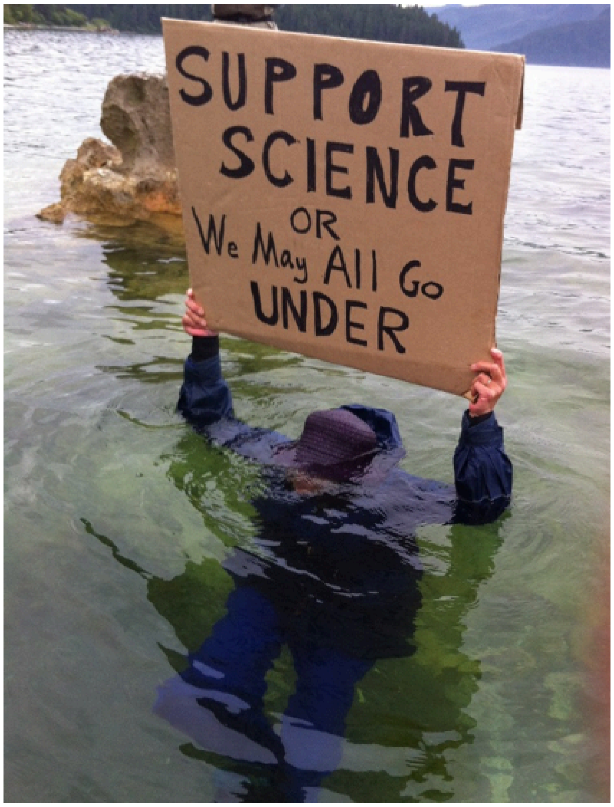
Sheila Watt-Cloutier, the former chair of the Inuit Circumpolar Council, wrote of the human rights claims of northern-latitude people: “We Inuit and other Northerners . . . are defending our right to culture, our right to lands traditionally used and occupied, our right to health, our right to physical security, our right to our own means of subsistence and our rights to residence and movement. And as our culture, again, as I say, is based on the cold, the ice and snow, we are in essence defending our right to cold.”

7. We must act, because personal integrity requires us to do what’s right.

When people are asked to rate their hope that humankind will find a way to maintain a livable climate — on a scale of one (not a snowball’s chance in hell) to ten (nothing to worry about) — they generally come in at about three to four on the hope-o-meter.[3] They speak wistfully: “Let’s face it. Our options are limited, our cities and homes and transportation systems are disgracefully designed, destructive ways of living are skillfully protected by tangles of profit and power around the world, extractive corporations are behaving like sociopaths (see “characteristics of a sociopath”), and we have run out of time. How can any reasonable person be hopeful? And if you don’t have hope, then why should you act?”

But to think there are only two options — hope and despair — is a fallacy of false dichotomy. Between hope and despair is the broad and essential expanse of moral ground, which is not acting out of hope or failing to act out of despair, but acting out of personal integrity.

3. Informal polling of the audience members who attend the climate ethics talks given by Kathleen Dean Moore and Michael P. Nelson, 2010-2017.



By Nick Olmsted, reprinted by permission.

Integrity: a matching between what you believe and what you do, which is walking the talk. To act justly because you believe in justice. To live gratefully because you believe life is a gift. To act lovingly toward the Earth, because you love it. The meaning of our lives is not in what we accomplish in the end, any more than the meaning of a baseball game is the last out. What makes our lives meaningful is the activities we engage in that embody our values, whatever happens in the world. What does integrity ask of us? First, to refuse to be made into instruments of destruction. With thoughtless decisions about what we invest in, what we buy, what we praise, what we value, what we do for a living, we volunteer to be the foot soldiers of corporate destruction. Soldiers used to say, "Hell no," to an unjust war. Can we say the same to an unjust, far more disastrous, way of life?

Integrity calls us to live in ways that express our deepest values. As we live with integrity, we can escape the unsettled grief of lives that violate our deeply held beliefs about right and wrong. As we live with integrity, we can imagine and bring into being new ways of living on the land that are bright with art and imagination, nested into families and communities, grateful and joyous – and lasting for a very long time.

Questions

1. If you were asked to rate your hope that humankind will find a way to maintain a livable climate on a scale of one (not a chance) to ten (nothing to worry about), what is your number on the hope-o-meter? Has it changed from the beginning of this class to the end? Why?
2. Of the seven reasons to take action to prevent climate catastrophe, which speaks most powerfully to you? To your classmates? (After you have made your decisions, it might interest you to compare it to the rankings of a prior Oregon State University class: 1. Because moral integrity requires us to do what is right. 2. Because justice demands it. 3. For the sake of all forms of life on the planet. 4. For the sake of the children. 5. To honor our duties of gratitude and reciprocity. 6. Because compassion requires it. 7. To protect the thriving of humankind – one lousy vote.)
3. Former President Obama argues that we are borrowing this planet from our children and our grandchildren. Play out that analogy. In what way, borrowing? What does that mean for how we should

act?
4. Suppose you come upon a person who is drowning and calling out for help. What effect does each of the following (considered separately) have on your personal responsibility to save the person?

    1. You know the person is drowning.
    2. You cannot swim.
    3. A thousand people are standing with you, watching the person flailing around.
    4. You didn't push the person into the water.
    5. You are busy, on your way to an important meeting.
    6. Saving the person would require some financial sacrifice, since you are wearing expensive clothes and watch, which would be ruined.Now, how is this case similar, and dissimilar from our personal responsibility to take action to counter climate change?

5. Australian philosopher Peter Singer argues that there is a fair and practical way to allocate the right to emit greenhouse gases: *Take the total capacity of the atmosphere to absorb greenhouse gases without harmful effects, divide it by the number of people on Earth. That is an individual's fair share. Allocate to each country an emissions quota equal to its population's shares. Then create a market in which countries that want a higher quota can buy shares from counties that emit less.* In your considered judgment, is this fair?
6. The Buddhist scholar Thich Nhat Hanh writes, "Our own life has to be our message." If so, then we had each better ask these questions: What message do I want my life to send? What might I do to send that message? What do I do now that might prompt someone to read the contrary message?

Videos

Lecture: Ethics

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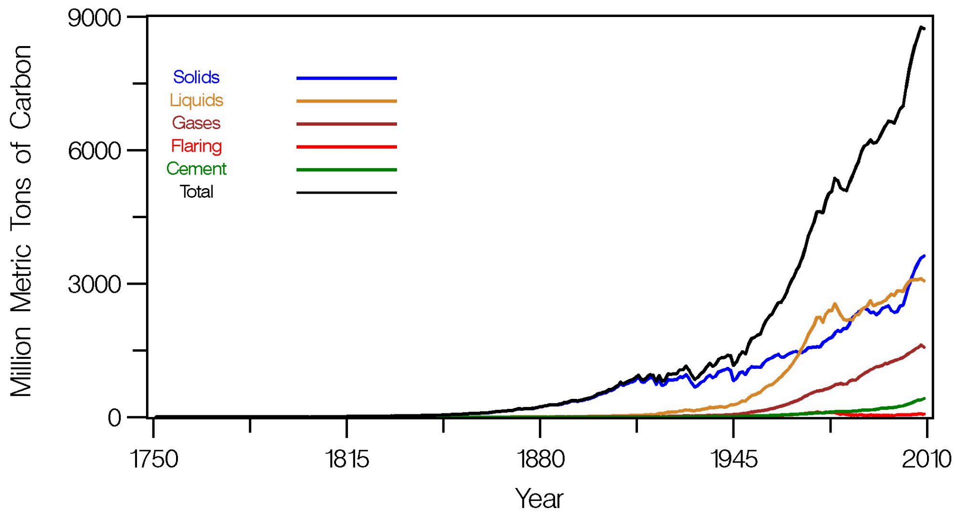
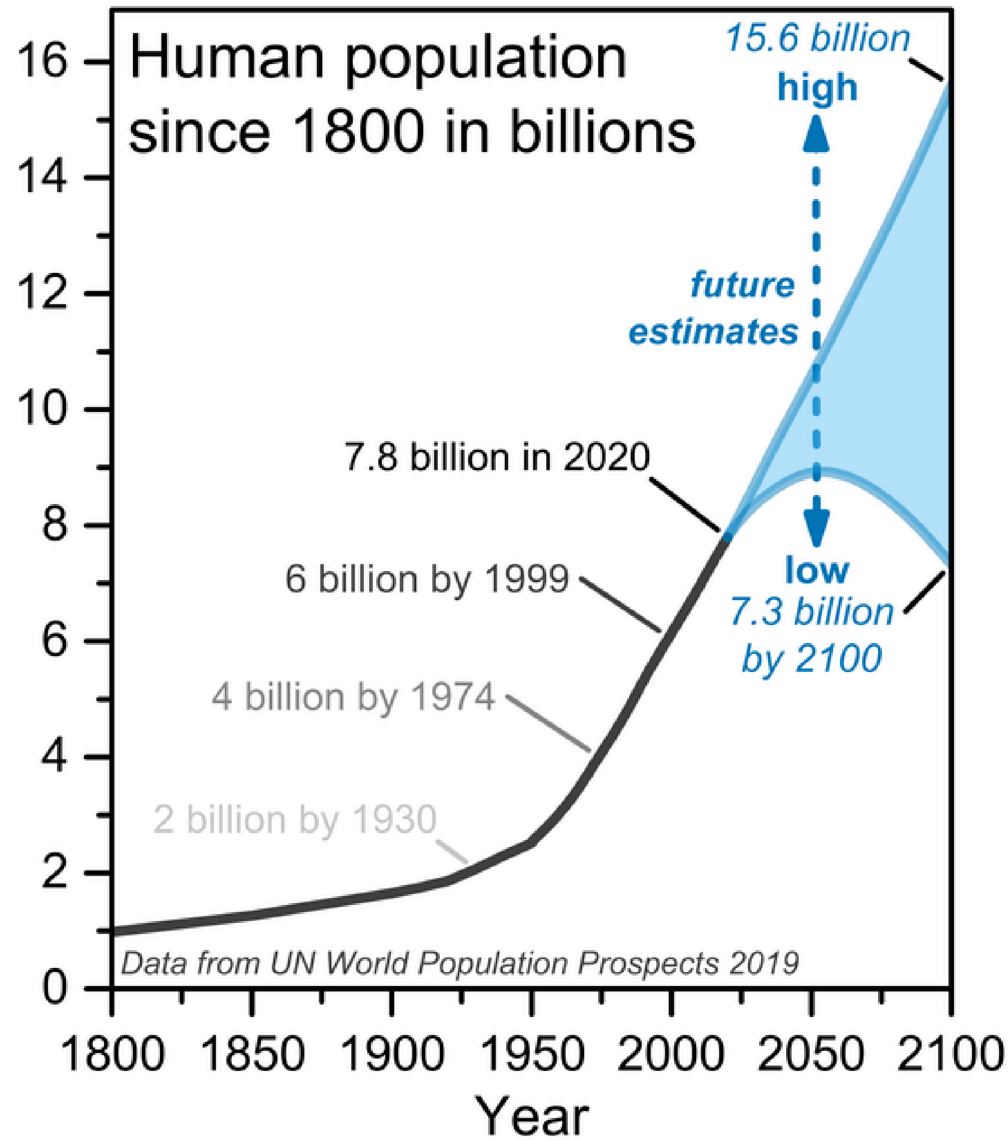
11. Solutions

Climate change is a difficult problem to solve because our modern society was build with fossil fuel burning as an energy source. We still depend strongly on fossil fuel energy for everything from driving our cars and washing our laundry to charging our cell phones and heating our homes. In order to stabilize climate, however, we'll need to move to near-zero carbon emissions in the long run. Thus, the challenge is to decarbonize our economy. The longer we wait with this transformation the larger the impacts of future climate change will be or the faster future emission reductions will need to be, if the goal is to stay within a certain limit of global warming. And because it is a global problem, the whole world, or at least most of it, will need to cooperate to solve it. Moreover, because we're already committed to further climate change, we better prepare to adapt to it. This chapter will discus technological options, political efforts and personal actions towards climate mitigation solutions and ways to adapt to the inevitable.

Fig. 1 illustrates that historically the increase in carbon emissions was caused by human population growth and an increase in the economy as expressed by e.g. the Gross Domestic Product (GDP). Both population and GDP per person have increased exponentially during the past ~200 years. The increased use of fossil-fuel-based energy has lifted many people out of poverty and improved the lives of millions although many people in the developing world remain in poverty. Formally, anthropogenic carbon emissions $ACE = P \times GDP/P \times E/GDP \times CE/E$ in a certain region or country can be decomposed into the population P, the GDP per person (GDP/P), the energy intensity of GDP (E/GDP) and the carbon emissions per energy (CE/E). Energy intensity of GDP has been decreasing during the past 50 years, which compensated somewhat the increase in population and GDP per person, whereas carbon intensity has not changed as much as the other factors (IPCC, 2014, Fig. SPM.3). This decomposition is known as the Kaya Identity. Currently the world population is almost 8 billion people and it is expected to continue to increase, at least for the near future. This increase will continue to put more pressures on the Earth system; climate change is just one of them. Others are, for example, increased occupation of wild places by humans, which reduces habitats for many species of plants and animals or increasing demand for resources such as food and fresh water. Population growth could be efficiently reduced by educating and empowering women in the developing world and by poverty reduction. Reducing the GDP per person is probably not a good way to reduce emissions, because most people do not want to reduce their standard of living and energy consumption (although some do and a lot of waste could also be cut without affecting our standard of living much). Since many people in the developing world hope to improve their standard of living it will be desirable to further increase the average GDP per person in the future. But if carbon emissions per energy could be reduced e.g. by shifting to non fossil fuel energy sources, that would reduce emissions without reducing GDP or energy consumption.


Million Metric Tons of Carbon
9000
6000
3000
0
Solids
Liquids
Gases
Flaring
Cement
Total
1750
1815
1880
1945
2010
Year


Human population
since 1800 in billions
15.6 billion
high
future
estimates
7.8 billion in 2020
6 billion by 1999
4 billion by 1974
2 billion by 1930
low
7.3 billion
by 2100
Data from UN World Population Prospects 2019
0
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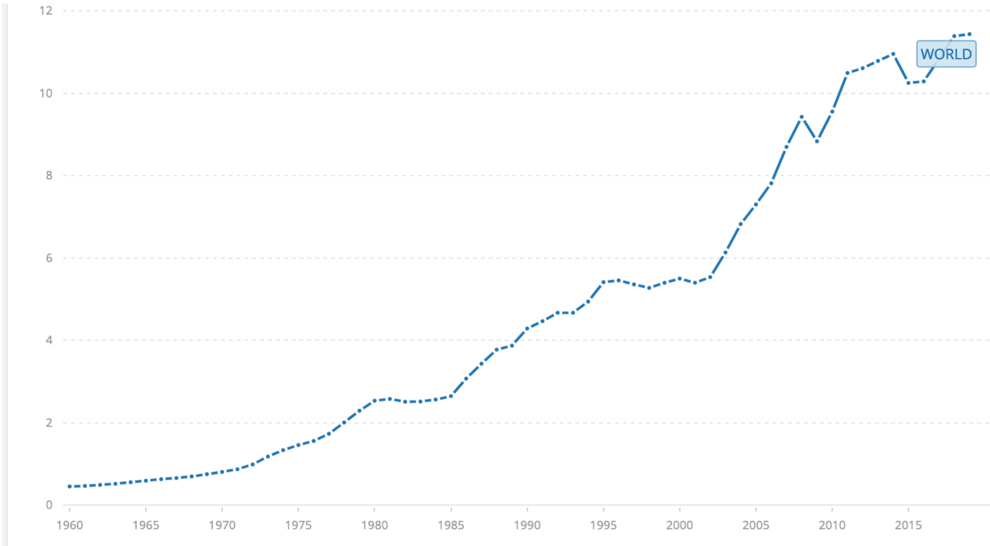

Figure 1: Evolution of global carbon emissions (top), world population (center) and per person GDP (bottom, in thousand US$). From cdiac.ess-dive.lbl.gov, wikipedia.org, and worldbank.org.

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a) Technology

Current global energy production relies heavily on fossil fuel burning (Fig. 2). The largest energy sources are oil, coal and natural gas, all of which are fossil fuels, whereas all non-fossil fuel sources together account for only about 20% of the total. Most oil consumption powers internal combustion engines in cars and trucks, which have a very low efficiency. Only about 25% of all energy input into transportation is used to move vehicles, whereas most of the energy is wasted as heat. Coal is mainly used in power plants to generate electricity, which is also associated with a loss of a little over one half. Note that this loss is less than that from internal combustion engines, which makes electric cars have lower carbon footprints than internal combustion engine cars even if the electricity is generated from coal. Most natural gas is used to generate electricity and to heat buildings. Hydro- and nuclear power are used exclusively to produce electricity, whereas biomass is used mostly for cooking and heating homes in the developing world. New renewables such as solar and wind supply only a small fraction of all energy.

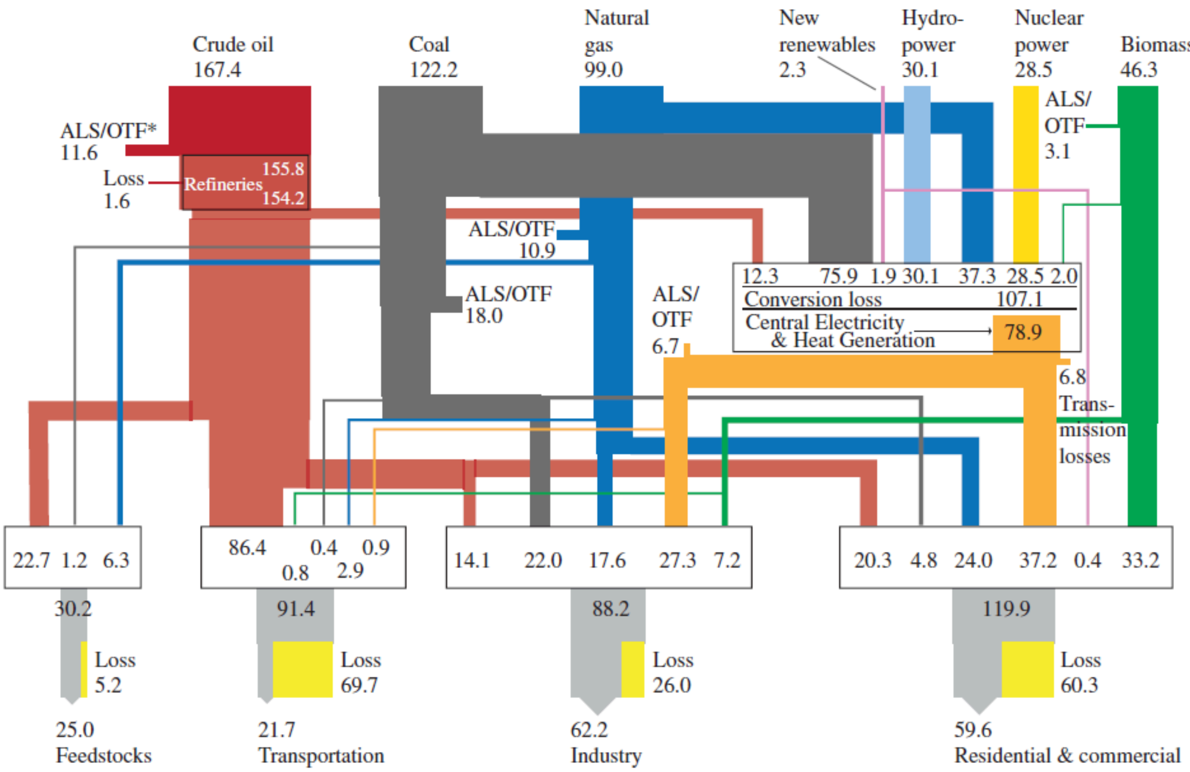

Figure 2: Energy flows in the world economy in 2005 in exajoules (1EJ = 10^{18}J) per year from primary sources top to end-use sector (bottom). ALS = Autoconsumption, losses, stock changes. OTF = other transformation to secondary fuels. From [GEA (2012)](). Click [here]() for a cool animation of the time-varying energy fluxes in the US.

However, in some countries **renewable** energy sources have seen a rapid increase in recent years. Germany, for example, has increased the renewables' contribution to total electricity production from 3% in 1990 to 40% in 2018 (45% in 2020), while its economy has been one of the strongest in Europe. I have witnessed this transformation during my visits as more and more solar panels appeared on rooftops and wind turbines in the fields. Denmark plans to move to 100% renewable energy by 2050. In the United States renewables currently account for 10% of total energy consumption and 15% of electricity production and it is rapidly increasing. In 2016, e.g., its solar power capacity doubled. The advantage of renewables is [almost unlimited potential supply]() with minimal carbon emissions (some emissions

Figure 3: Photograph of Solar Panels and Wind Turbines.

occur during the production and installation of solar panels and wind turbines) and that they are not associated with the dangers of nuclear power. Their disadvantages used to be their cost, particularly their high up-front investment cost. Once installed, however, solar panels and wind turbines operate almost without maintenance cost since solar energy and wind is free. During the past 10 years cost for solar panels have decreased dramatically by 80%. Thus, if viewed over the lifetime of a system renewables become competitive with fossil fuels. Other renewable energy sources are geothermal heat, tide and wave energy and hydroelectric dams. Another issue with renewables is their intermittent energy supply. Solar panels only work during the day, whereas wind turbines only work when the wind blows. However, a recent study showed that 80% of all electricity demand could easily be covered by wind and solar (Shaner et al., 2018).

Figure 4: Photograph of Workers at the Damaged Fukushima Nuclear Reactor.

Nuclear plants also provide power that is fossil-free and does not cause carbon emissions (except during construction). For this reason they are viewed by some as an important future energy source. However, nuclear power has disadvantages too. Not only are they expensive to build they are also dangerous to operate and they produce radioactive waste for which currently no long-term repository exists. Catastrophic accidents such as the nuclear meltdowns in 2011 at the Japanese Fukushima Daiichi plant and in 1986 at the Chernobyl reactor in what is now northern Ukraine, have shown the dangers associated with nuclear power production. Thus, nuclear power remains a controversial topic.

Figure 5: Photograph of an Electric Car.

Currently many companies move towards more **electric cars** or hybrid vehicles. Due to their much higher efficiencies (80-90%) their energy use is much smaller than for cars with internal combustion engines and their carbon footprint can be close to zero (some is associated with vehicle production) if renewable energy sources are used for the electricity. Even if fossil fuels are used for the electricity the carbon footprint of electric cars is still lower than that of internal combustion engine vehicles. Considering the large amounts of carbon emissions that currently come from the transportation sector and the large losses that occur there (Fig. 2) shifting to an electric vehicle fleet could bring a tremendous reduction in future emissions. Even though electric cars are more expensive to purchase their lifetime costs are lower than gasoline powered cars. This is because the average cost for equivalent electricity ($1.10) is less than half the cost for a gallon of gas ($2.50). Electric cars have other advantages too.

No oil change, no pollution, more torque. Manufacturing electric cars, however, is not without environmental or human impacts, such as from the mining of raw materials used in the production of batteries.

Increasing energy efficiency is another cost-effective way to reduce carbon emissions. Fig. 2 shows that residential and commercial buildings waste about half of their energy. Building insulation does not only lower its carbon emissions but it also saves the owner money. When we bought our house it was not insulated. Adding insulation reduced our heating costs and made the house more comfortable to live in. It made it quieter too. A few years ago we also replaced our old conventional electrical hot water heater with an inexpensive new heat pump water heater. This simple change reduced our electricity bill by half.

Carbon capture and storage has been proposed and tested to reduce carbon emissions from fossil fuel powered plants or to remove carbon from the atmosphere. However, this technology is expensive, not well developed, and the storage in geological formations may be leaky. Here is an interesting video about natural and artifical ways to remove carbon from the air and store it in the ground.

b) Politics

In 1992 the United Nations Framework Convention on Climate Change (UNFCCC) was adopted at the Rio Earth Summit. It was ratified by 197 countries (Parties to the Convention). Its Article 2 states:

> *The ultimate objective of this Convention and any related legal instruments that the Conference of the Parties may adopt is to achieve, in accordance with the relevant provisions of the Convention, stabilization of greenhouse gas concentrations in the atmosphere at a level that would* ***prevent dangerous anthropogenic interference with the climate system****. Such a level should be achieved within a time frame sufficient to allow ecosystems to adapt naturally to climate change, to ensure that food production is not threatened and to enable economic development to proceed in a sustainable manner.*

The UNFCCC recognizes that the developed world is responsible for most historical carbon emissions and thus it should lead the way to reduce its emissions. The Kyoto Protocol, an international agreement linked to the UNFCCC, commits its Parties, which are developed countries, to binding carbon emission reduction targets. However, it was not successful in reducing global carbon emissions, presumably at least in part because the United States has never ratified it, Canada has exited it and Russia has not agreed to emission reductions in its second phase (2013-2020). The US under the administration of President George W. Bush argued that it was not fair to reduce US emissions, while China was allowed to increase its emissions, even though per capita emissions in the US were much higher than those in China. The Paris Agreement has the goal to limit global warming to well below 2°C and it was ratified by 174 countries. It has made emission reductions essentially voluntary by soliciting pledges from each country and includes developing countries like China. However, resistance against any reduction in carbon emissions was evident when the US administration under President Trump withdraw from the agreement although the current Biden administration has re-entered the agreement.

c) Personal Actions

Each of us can take individual actions to reduce carbon emissions and the impacts of climate change. Conserving energy e.g. by switching off lights when they are not needed. Switching to energy efficient appliances such as LED lights. Recycling also helps. Driving less or choosing an energy efficient car like a hybrid or an electrical car or a smaller vehicle will reduce your carbon footprint. Even better is riding your bike. I like to ride my bike not only because I want to reduce carbon emissions but also because it is fun and it is healthy. It's my daily exercise. Not flying so much also helps. I use skype to communicate with my family in Germany. Eating less meat is another way to reduce your carbon footprint because meat production requires more carbon emissions than vegetarian food. I'm not a vegetarian myself but I like tasty vegetarian food. Homeowners can weatherize their homes or install solar panels. We've installed 20 solar panels in 2017. The system produces enough electricity for all of our family needs including that for an electric car. There are many things we can do as individuals. You can check your own carbon footprint at various websites such as this to help you with this process.

Ultimately, however, individual actions have limited effects. Imagine, for example, that you start using less gas by riding your bike instead of driving your car. If many people would do it this would reduce demand for gas and it would make gas cheaper, assuming supply stays constant. Cheaper gas has the effect that other people buy larger cars and use more gas, thus diminishing or erasing the effects of your personal emission reduction. I'm not saying that we should not take individual actions. To the contrary. I think it is important that everybody does her/his part in making the world a better place and that we act in accordance with our believes. However, we should not have delusions as to the effects of individual actions. Therefore I believe collective action is also needed. In my opinion only governmental actions such as a carbon tax or a cap-and-trade system will be able to accelerate the transition to renewable energy sources that we need to avoid potentially dangerous climate change. Thus, perhaps the most important individual actions we can take is to vote and to engage in the political process to make change happen. Part of that is talking to family, friends, neighbors and politicians.

d) Adaptation

Climate change does already affect us. It is not only a thing of the future or of far-away places. It is an issue of the here and now. Since we cannot avoid further climate change in the immediate future we will have to adapt to it if we don't want to suffer its consequences. Perhaps one of the most important things is to be aware that it is happening. This will make us think about possible consequences for the region we live in. If regional projections are available we can consider them in planning. If not we can still extrapolate past changes according to the principle that we can expect more of what we've observed in the recent past. In Oregon, for example, we can expect less snow pack and lower summer stream flows. Eastern Oregon, which is already dry, can expect even dryer conditions. Along the coast sea levels will continue to rise and the ocean will continue to acidify. Regionally changes in sea level or ocean acidification may be different from global mean changes.

Questions

- What is mitigation of climate change?
- What is adaptation to climate change?
- What are the two main reasons for the increase in historical carbon emissions?
- What is the Kaya Identity?
- Considering the Kaya Identity, what could be done to reduce carbon emissions in the future?
- List five things that can be done to mitigate climate change.
- What is the goal of the United Nations Framework Convention on Climate Change as expressed in its Article 2?
- What personal actions would you consider to take in order to mitigate climate change impacts?
- Why will we need to adapt to future climate change impacts?
- Think about a particular region/country of the world that may be affected by climate change. What measures could be taken to adapt to the changes?

Videos

[Lecture: Solutions](#)
[Lecture: Politics](#)

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Glossary

Absorption

Uptake of something. E.g. a photon of electromagnetic radiation is absorbed by a molecule. Earth's atmosphere absorbs most infrared radiation from the surface.

Adaptation

changing our ways of living in order to cope with the impacts of climate change. E.g. building dikes to avoid flooding from sea level rise.

Aerosols

Small particles in the air. Aerosols reflect sunlight back to space and therefore lead to cooling of the surface (negative forcing). Aerosols originate from natural (dust, ash from volcanic eruptions, wave breaking) and anthropogenic (smoke) sources.

Albedo

Reflectivity. Snow, ice, clouds and other bright surfaces have a high albedo, which leads to most sunlight being reflected to space. Ocean and vegetation on land have low albedos. They absorb lots of the suns radiation, which leads to warming.

Allowances

Authorizes a utility or industrial source to emit emissions during a given compliance period. Allowances are fully marketable commodities. Once allocated, allowances may be bought or sold in a trading market.

Altitude

Height.

Anomaly

An anomaly is a difference or change with respect to a reference.

Anthropogenic

Human made. Anthropogenic climate change is the change that can be attributed to human activities. Humans cause climate change by emitting greenhouse gases and aerosols, and by land use change.

Archives

The material from which paleoclimate proxies are obtained, e.g. tree-rings, ice-cores, ocean sediment.

Attribution

The causes (natural and/or anthropogenic) of recent climate change.

Benthic

Living on or in ocean sediments.

Biological Pump

The removal of carbon from the surface and sequestration in the deep ocean by marine biota. Phytoplankton take up carbon during photosynthesis at the sunlit surface. They are eaten by zooplankton and the organic matter is transferred through the food web to higher trophic levels. Some of the organic matter sinks to depths, where it is remineralized by bacteria.

Biome

a community of plants and animals occupying a major habitat

Blackbody

A body that is able to absorb (and emit) electomagetic radiation at all frequencies. The Stefan-Boltzmann law states that the total energy flux from a blackbody is proportional to the fourth power of its temperature Fblackbody = σT4, where σ =5.67×10-8W/(m2K4) is the Stefan-Boltzmann constant and T is the temperature in Kelvin.

Boundary Conditions

Boundary conditions are values for prognostic variables (e.g. temperature) or fluxes (e.g. heat flux) at the boundary of the model domain needed to solve the interior grid boxes in climate models and other models that use differential equations.

Budget Equation

Rate-of-Change = input – output. A budget in balance (or equilibrium) does not change: input = output.

Chronology

Assigning time to a paleoclimate record.

Climate Sensitivity

$S_c = \Delta T/\Delta F$ is the global average surface temperature change ΔT divided by the forcing ΔF. Units are K/(W/m2). Also, commonly used is the temperature change for a doubling of CO_2, $\Delta T_{2xCO2} = S^*\Delta F_{2xCO2}$, which can be converted easily because we know ΔF_{2xCO2} = 3.7 W/m2. ΔT_{2xCO2} is not very well known. It is most likely somewhere between 1.5 and 4.5 K. Most studies suggest it is about 3 K.

Climatology

Average seasonal cycle.

Condensation

Transition of a substance (e.g. water) from vapor to liquid. Condensation occurs when air is saturated with water vapor and condensation nuclei (e.g. small particles) are present. Latent heat is released during condensation.

Convection

Vertical overturning of air or water due to unstable stratification. If air is heated at the surface it can lead to upward motion in the atmosphere. In the ocean convection can occur when surface waters are cooled. Deep convection in the ocean only occurs at high latitudes where surface waters are close to the freezing point.

Convergence

Flow towards a point. Convergence of surface waters in the ocean lead to downwelling. Convergence of surface winds near the equator lead to rising air in the intertropical convergence zone. See also divergence.

Corliolis Force

deflection of moving air or water masses towards the right (left) in the northern (southern) hemisphere due to the rotation of the Earth.

Delta

Greek capital letter (Δ) usually used to indicate a change or difference in something. E.g. $\Delta T = T_2 - T_1$ temperature change. A small delta (δ) is used for isotopes and this version (∂) indicate differentials (infinitesimally small changes). E.g. $\partial T/\partial t$ is the rate of change of temperature at any given time. It can be approximated by a difference $\partial T/\partial t \cong \Delta T/\Delta t$.

Demand

Describes a consumer's desire and willingness to pay a price for a specific good or service.

Density

Mass divided by volume. The density of air is approximately ρair = 1 kg/m3 (greek letter rho) and that of water is ρwater = 1,000 kg/m3. Density of air and water is lower if it is warmer. Sea water is also more (less) dense the saltier (fresher) it is.

Detection

The question whether recent observed climate change is significant, that is outside the range expected from natural internal fluctuations only.

Divergence

flow away from one point. See also convergence.

EBM

See models.

Eccentricity

Earth's orbit around the sun is not a perfect circle but it is slightly elliptical (egg shaped). The degree of deviation from a perfect circle varies on with ~100,000 year cyclicity and is called eccentricity.

El Niño

El Niño years are warmer than usual temperatures in the central and eastern tropical Pacific associated with a shift of atmospheric convection from the West Pacific warm pool towards the center and eastern Pacific. La Niña years are the opposite (colder than usual temperatures in the eastern Pacific). El Niño is a phenomenon of natural, internal (unforced), variability of the coupled ocean-atmosphere system with a typical timescale of 3-7 years. It is also known as El Niño Southern Oscillation (ENSO).

Emission

Production and discharge of something. E.g. Earth's surface emits longwave radiation. Humans emit carbon into the atmosphere.

Emission standard

A standard that sets a quantitative limit on the amount of a specific air pollutant that may be released from a specific source over a given timeframe.

Energy

In physics, energy E comes in different forms. Mechanical energy E = W is equal to the work W = F*d done by displacing an object a distance d with the force F. Force F = m*a, with units of kg*m/s^2 = N (newton), is mass m times acceleration, where acceleration a = dv/dt is the change in velocity v per time t. Thus the unit of energy is kg*m^2/s^2 = J (joule). Power P = W/t is energy per time and has units of watts (1 W = kg*m^2/s^3). In thermodynamics the internal energy or heat content E = C*T of an object is proportional to its temperature T, where the constant C is the heat capacity.

Evaporation

Transition of a substance (e.g. water) from liquid to vapor phase. The rate of evaporation from the ocean depends on sea surface temperature (the warmer the more evaporation), the relative humidity of the air (the drier the air the more evaporation), and the wind velocity (the more wind the more evaporation). The energy required for that transition is called the latent heat of vaporization.

Externality

The cost (or benefit) that affects a party who did not choose to incur that cost (or benefit).

Extreme Events

Rare weather or climate events such as hurricanes, typhoons, floods, droughts, and tornadoes that can often be damaging.

Fallacy of False Dichotomy

A mistake in reasoning, in which one assumes there are only two alternatives, whereas there are in fact more.

Feedback

A change in the climate system as a response to a radiative forcing that will amplify (positive feedback) or dampen (negative feedback) the initial forcing. E.g. initial forcing of increasing CO2 leads to warming, which leads to more evaporation and water vapor in the air, which leads to more warming (because water vapor is a greenhouse gas). Important feedbacks are the Planck (negative), water vapor (positive), ice-albedo (positive), lapse rate (negative) and cloud (positive or negative) feedback. The sum of all feedbacks determines the climate sensitivity.

Fluxes

Amount of energy or matter flowing through an area for a certain time. E.g. the global average incident solar radiation energy flux S=342 W/m2 is the amount of energy in units of joules (J) flowing through an area of one square meter (m2) in one second (s). Watt (W) is the unit of power (1W = 1J/s). A 100 W incandescent light bulb is quite bright. Three and a-half of those per square meter is about the same energy flux as S.

Foraminifera

Microscopic zooplankton. Planktic foraminifera live near the surface, benthic foraminifera live on or in the sediments. Foraminifera build shells out of calcium carbonate (CaCO3) that record changes in their environment. Measurements on fossil shells are important paleoclimate proxies.

Forcing

Radiative forcing (ΔF) is a change in energy fluxes F (in W/m2) at the top-of-the-atmosphere that causes climate change. It is defined as positive (negative) if it leads to warming (cooling). The radiative forcing for a doubling of CO2 is ΔF2xCO2 = 3.7 W/m2. Other examples are increased solar radiation (positive), increased aerosols (negative) or increased surface albedo (negative), e.g. due to land use changes.

Fractionation

Processes fractionate isotopes when one isotope (e.g. 16O) is preferred with respect to another isotope (18O). Evaporation, for example, favors water molecules with 16O, which causes more light water (H216O) to evaporate compared to the heavier H218O. Thus, the ratio R=18O/16O (and hence the δ18O) of water in the vapor phase will be smaller than that in the liquid phase. During condensation, the opposite happens and the heavier isotopes are preferred, which makes the cloud droplets have a higher δ18O value than the vapor they form from. Carbon isotopes fractionate during photosynthesis, such that plants (including algae) preferentially use the light 12C.

GCM

See models.

Great Turning

A phrase coined by visionary Joana Macy, "the transition from a doomed economy of industrial growth to a life-sustaining society committed to the recovery of our world."

Greenhouse Effect

Greenhouse gases in the atmosphere absorb part of the longwave radiation emitted from the surface of the Earth. They re-emit this radiation in all directions, half of which goes back down towards the surface, which leads to warming there. The natural greenhouse effect leads to 33 K (33°C; 60°F) warmer surface temperatures on Earth. Without greenhouse gases Earths surface would be frozen. The enhanced greenhouse effect is due to increases in greenhouse gases due to human activities and leads to global warming.

Greenhouse Gas

Greenhouse gases are molecules that are able to absorb and emit electromagnetic radiation in the infrared part of the spectrum (longwave). Water vapor (H2O), carbon dioxide (CO2), methane (CH4), and nitrous oxide (N2O) are Earth's most important greenhouse gases.

Hadley Cell

Atmospheric circulation in the tropics characterized by rising motion near the equator in the Intertropical Convergence Zone, poleward motion aloft (~10 km), sinking in the subtropics, and equatorward flow at the surface (trade winds).

Heat Capacity

The amount of heat required to increase the temperature of a substance by one degree Celsius. The specific heat capacity of air at constant pressure cp = 1 J/g°C. That of water is 4.2 J/g°C.

Holocene

The last 10,000 years of Earth's history. A warm and stable climate. The early Holocene (10,000 – 5,000 years before the present) was about as warm as the last 30 years. After that climate cooled slowly into the Little Ice Age (1,500 – 1,800 AD).

Human rights

The shared standards of achievement for all people and nations, expressed in 1948 in the Universal Declaration of Human Rights, which begins with the right to life, liberty, and security of person.

Hydrological Cycle

The water cycle.

Insolation

Incoming solar radiation.

Intertropical Convergence Zone

(ITCZ) convergence of surface winds near the equator leads to rising air, cloud formation and precipitation. The ITCZ constitutes the upward branch of the Hadley Cell.

Intrinsic Value

As opposed to instrumental value (or use value, or value as a means to some other end) something is said to possess intrinsic value if it has value in and of itself, quite beyond or apart from the use to which it might be put.

Inuit

A group of culturally similar indigenous peoples inhabiting the Arctic regions of Greenland, Canada and Alaska

IPCC

Intergovernmental Panel on Climate Change: an international effort to synthesize the most recent science on climate change. The IPCC's assessment reports are published every 6 or 7 years and contain a comprehensive summary of the peer-reviewed scientific literature including the most recent projections.

Irradiance

Flux of electromagnetic radiation (energy) through a surface per unit area. Units are watts per square meter (Wm-2).

Isotopes

Versions of the same element with a different mass (different number of neutrons). Usually there is one common (abundant) isotope and one or more rare isotopes. E.g. carbon exists as carbon 12 (regular carbon), carbon 13 (one more neutron than carbon 12) and carbon 14 (radiocarbon, two more neutrons than carbon 12). Radiocarbon is radioactive, which means that it is unstable and decays with a half-life of about 5600 years. Isotopes are often reported in delta values, e.g. δ13C = (R/Rstd-1)*1000 for carbon 13, which have units of permil (‰), where R=13C/12C is the ratio of carbon 13 over carbon 12 of the sample and Rstd is the ratio of a reference standard. Isotope delta values can change due to fractionation.

Jet Stream

Fast westerly winds in the upper atmosphere at subtropical latitudes, caused by the Coriolis force acting on the upper branch of the Hadley circulation.

Joules

Joule is the SI unit of energy or work. 1 J = 1 kg m2 s-2.

Kinetic Energy

Energy due to movement of an object. The kinetic energy of an object is equal to one half its mass times its velocity squared.

Land Use Change

Humans' effects on the climate system through modifications of the land surface e.g. through deforestation and agriculture.

Land-Sea Contrast

Temperature changes over land are usually larger than over the ocean. There are two main reasons for this: 1) in a transient situation (non-equilibrium) the larger heat capacity of the ocean delays ocean temperature changes compared to land, and 2) evaporative cooling is limited by the availability of water on land, whereas it is not limited over the ocean.

Lapse Rate

Rate of temperature decrease with height in the troposphere ΔT/Δz. On average the lapse rate is ~6.5 K/km, which is close to the moist adiabatic lapse rate. A dry atmosphere would have a lapse rate of ~10 K/km.

Last Glacial Maximum

The height of the last ice age, about 20,000 years ago. Large ice sheets covered parts of North America and northern Europe, sea-level was 120 m lower than it is today, the air was dustier, and the vegetation distribution was much different from today in many regions.

Latent Heat

Energy required for a phase change. E.g. to evaporate 1 g of water 2,300 J is required. The same amount of energy is released during condensation.

Marginal Benefit

The additional satisfaction or utility that a person receives from consuming an additional unit of a good or service. A person's marginal benefit is the maximum amount he is willing to pay to consume that additional unit of a good or service.

Marginal Private Cost

The private cost of an additional unit of output of a good experienced by an individual firm. Does not include external costs (the social or environmental costs which may arise from the production of a good).

Marginal Social Cost

The total cost society pays for the production of another unit. The total cost of the production includes costs to others and the environment as a whole. MSC is calculated as marginal private cost plus marginal external cost (the cost of externalities).

Meridional

North-South.

Meridional Overturning Circulation

Deep ocean circulation driven mostly by density differences.

Mitigation

reducing carbon emissions in order to lessen their negative impacts

Models

Climate models are based on conservation equations for energy, mass, momentum, water, salt, carbon, and other substances. The simplest climate model is the zero-dimensional (0D) Energy Balance Model (EBM), which solves just one equation for the global average surface temperature. Slightly more complex models are 1D EBMs and 1D (vertical) radiative convective models. Intermediate complexity models are e.g. 2D EBMs. The most comprehensive are General Circulation Models (GCMs).

Obliquity

Tilt in Earth's axis with respect to Earth's orbit around the sun. Varies at ~40,000 year cycle.

Opaque

Not transparent. A medium is opaque if radiation cannot pass readily through it. Earth's atmosphere is opaque to terrestrial radiation.

Ozone

$O3$ is a radiatively active molecule in Earth's atmosphere. It occurs mainly in the stratosphere where it absorbs ultraviolet radiation.

Ozone hole

Lack of $O3$, particularly over the Antarctic, caused by anthropogenic emissions of chlorofluorocarbons (CFCs) that were used as cooling agents in refrigerators and in spray cans.

Paleoclimate

Changes in climate before the instrumental record began ~100 years ago.

permil or ppt

parts per thousand (‰).

Photosynthesis

Process by which plants (including algae) turn $CO2$ and water into organic matter and oxygen using light as the energy source. It is the reverse chemical reaction of respiration.

Polar Amplification

Climate changes at high latitudes are larger than at low latitudes. One reason for this is the ice-albedo feedback, which amplifies climate changes at the poles. Another reason is more latent heat transport from the tropics towards higher latitudes in a warmer climate.

Polar amplification

Pollution

Substance with harmful effects put into the environment by humans.

ppb

Parts per billion.

ppm

Parts per million.

Precession

Wobble of Earth's axis with respect to its orbit around the sun. Varies at ~20,000 year cycles.

Precipitation

rain, hail or snowfall

Premise

A statement of fact or value offered in support of a conclusion – we also might think of this as "evidence" for a given position.

Prescriptive

A prescriptive statement is one suggesting we ought or ought not do something, as opposed to a descriptive statement that merely describes the facts of the matter.

Projections

Projections are predictions of future climate changes assuming specific scenarios for future anthropogenic radiative forcings such as the emissions or concentrations of greenhouse gases.

Proxies

Surrogates for climate variables used in paleoclimate research. E.g. pollen can be used to reconstruct past vegetation cover, which allows inferences on temperature and precipitation.

Radiation

Electromagnetic radiation are waves of electric and magnetic fields that travel through vacuum (space) and matter (atmosphere, ocean). Important for Earth's climate is radiation from the sun, also called solar, or shortwave radiation FSW, which is at shorter wavelengths (in the visible, ultraviolet and near infrared parts of the spectrum) than the radiation emitted from Earth. The latter is called terrestrial, thermal, or longwave radiation (FLW).

Radiative Forcing

(ΔF) Changes in energy fluxes F (in W/m2) at the top-of-the-atmosphere that cause climate change. It is defined as positive (negative) if it leads to warming (cooling). The radiative forcing for a doubling of CO2 is ΔF2xCO2 = 3.7 W/m2. Other examples are increased solar radiation (positive), increased aerosols (negative) or increased surface albedo (negative), e.g. due to land use changes.

Reflection

Bounces off something.

Relative humidity

(*rh* = *q*/*q*sat) the amount of water in the air (specific humidity *q*) relative to its temperature dependent saturation value *q*sat. It is typically reported in percent. E.g. if air at 30°C has a specific humidity of *q* = 15 g/kg its relative humidity is approximately *rh* = 15/30 = 50% since the saturation specific humidity is ~30 g/kg (see Fig. 16 in Chapter 4).

Resolution

Coarse resolution means that details are not apparent, whereas fine resolution depicts more details, both in space and time. E.g. a coarse resolution climate model does not represent spatial details of the real world. A high-resolution paleoclimate record can depict details in time of climate variations at a certain location.

Respiration

Bacteria, fungi, animals, and humans respire organic carbon by oxidizing it. This releases energy and produces water and carbon dioxide. It is the reverse chemical reaction of photosynthesis.

Salinity

The amount in grams (g) of salt per kilogram (kg) of sea water. Typical values in the open ocean are around 30-40 g/kg. Other often used and equivalent units are permil (‰) and practical salinity units (PSU): 1 g/kg = 1 ‰ = 1 PSU. Salinity influences the density of sea water such that it becomes denser the saltier it is.

Sea ice

Frozen sea water that swims of the ocean's surface. During the freezing process, much of the salt originally contained in the sea water is trapped in brain pockets and eventually lost by flowing slowly down through channels into the underlying water. Therefore, sea ice is almost fresh water with a *salinity* of only about 5 permil.

Sensible Heat

Heat can be transferred by an air or water parcel with a warmer temperature moving to a place where it is colder. This is called sensible heat flux.

Sociopath

Characterized by a lack of regard for the moral or legal standards in the local culture. There is a marked inability to get along with others or abide by societal rules.

Sound Conclusions

The conclusion of an argument that both follows directly from the premises or evidence provided, and that has only true premises.

Specific humidity

the amount of water vapor (in grams) per kilogram of moist air.

Standard Deviation

In statistics the standard deviation is a measure of variability around the mean of some data. To calculate it take the difference of all data from the mean, square it, sum it up, divide the sum by the number of data, and take the square root.

Stefan-Boltzmann Law

See Blackbody

Stratification

Layering of the ocean or atmosphere according to density. Stable stratification has light fluid on top of heavy fluid. If heavy fluid is on top of light fluid the stratification is unstable and convection will occur.

Stratosphere

Region of Earth's atmosphere, between about 10 and 50 km altitude, where temperatures increase with height. Holds about 20% of Earth's atmospheric mass.

T (Temperature)

Temperature. International units are Kelvin (K) and degrees Celsius (centigrade). In the U.S. Fahrenheit (F) is used. To convert use T[°F] = T[°C]*9/5 + 32 and T[°C] = T[K] - 273.15

t (Time)

Time (units are seconds).

Thermohaline Circulation

(aka Meridional Overturning Circulation) Density driven (deep) ocean circulation. Temperature (thermo) and salinity (haline) of sea water determine its density. Waters sink from the surface to the deep ocean only at a few places (in the North Atlantic and in the Southern Ocean around Antarctica) where it is dense enough. From there they flow into the interior of the oceans forming a global system of currents.

Trade Winds

Easterly (blowing from east to west) surface winds in the tropics.

Transmit

Passing through something. E.g. most shortwave radiation is transmitted through the atmosphere, whereas most longwave radiation is *absorbed* by the atmosphere.

Transparent

A medium is transparent if radiation passes through it. E.g. Earth's atmosphere is mostly transparent to solar radiation.

Tropopause

Region in Earth's atmosphere between the troposphere and stratosphere where temperatures do not change with height.

Troposphere

Lower (~10 km) part of Earth's atmosphere. Temperatures decrease with height in the troposphere. The troposphere includes most (about 2/3) of the mass of the entire atmosphere.

Worldviews

Answers to the most basic questions in philosophy – what is this world, what are humans, what is the relationship between this world and humans – constitute a person's or a group's worldview.

Zonal

East-West.

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This page provides a record of changes made to this guide. Each set of edits is acknowledged with a 0.01 increase in the version number. The exported files for this toolkit reflect the most recent version.

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Version	Date	Change Made	Location in text
0.1	MM/DD/YYYY		
0.11	10/02/2020	Links to external sources updated	All
0.12	03/01/2021	Updated Fig. 6	Chapter 2
0.13	04/26/2021	Various figures updated; added links to lecture videos	All