Climate Physics Chapter 1: Introduction

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An overview of some of the issues considered in the study of climate physics.

1 What is Climate?

When we think of climate, we usually think of temperature, atmosphere, wind and rain. In fact there are a number of factors that control and regulate Earth's climate, which has a very narrow temperature range compared to other planets. Some of these factors involve the radiation Earth absorbs and give off, the atmospheric composition, including the amount of greenhouse gases in the air, and the presence of clouds.

2 Faint Young Sun

Ever since early Earth cooled enough to form a solid crust 4.4 billion years ago, the Sun's radiation has been our main source of surface heat. Although Earth's temperature has experienced fluctuations, such as ice ages, we have experienced relative stability over time. We must not have had surface temperature fluctuations that were too large, or life would not have been able to evolve.
However, there are many factors that could cause changes in climate. Changes in the atmosphere affect the ability of radiation to enter or leave Earth. Atmospheric change can be due to bombardments that carry gases in or splash it away, volcanos that spew gases into the sky, random escape of gases due to thermal fluctuations, solar winds that blow away the outer atmosphere, chemical reactions that cause atmospheric elements to be bound in rocks, and even the presence of life. Bacteria can create methane, releasing it into the air, or pull nitrogen out of the air and fix it in the ground. Other forms of life introduce oxygen into the air via respiration, and chemical reactions with the oxygen can radically change the atmosphere.
Changes in the Sun's luminosity also affect Earth's climate. As the sun ages, it gets brighter, according to the formula:

$$\mathcal{L}(t) = \frac{\mathcal{L}(t_s)}{1 + \frac{2}{5}(1 - \frac{t}{t_s})} \tag{1}$$

Where $L(t)$ is the luminosity of the sun at time t, and t_s is the age of the sun.
This means that in the past, Earth received less solar radiation than it does now, yet it still had roughly the same climate. For this to occur, the atmosphere must have some self-regulating mechanism to counterbalance the sun, so that as the sun brightened over time the Earth's temperature remained fairly constant.

3 Goldidilocks Planet: Earth, Mars, and Venus

We have some examples of varying climate within our own solar system.
In Venus, we have a very hot planet. It has a temperature of 737 K. Sulfur dioxide clouds reflect away almost all of the incoming sunlight, but a thick atmosphere of CO_2 keeps in the heat. It also keeps the surface pressure at a level 92 times larger than that of Earth. Presumably, Venus started somewhat like Earth,

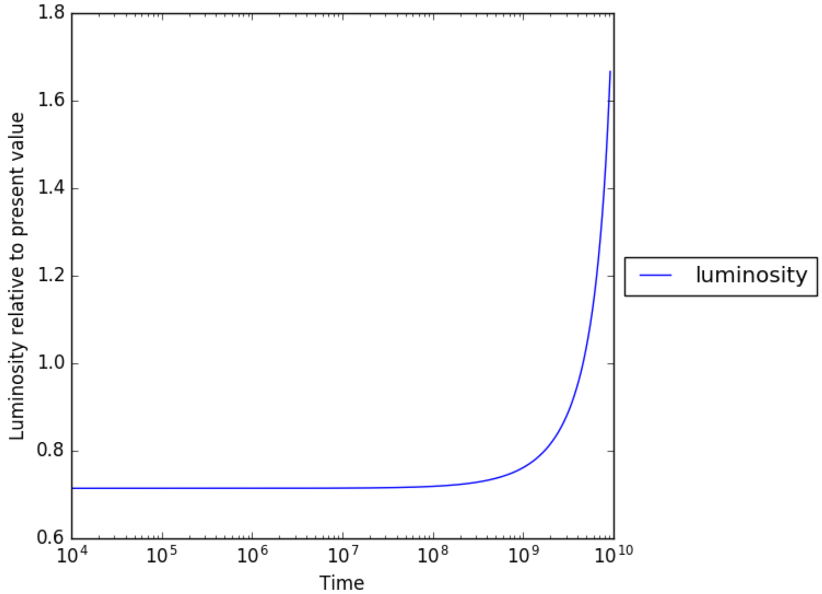

Figure 1: The luminosity of the sun over time. The current age of the sun is around $4.6 \cdot 10^9$ years. The graph is normalized to show a luminosity of 1 at the sun's current age.

but due to a runaway greenhouse effect, where the surface became warm enough to evaporate water, which acted as a greenhouse gas in the atmosphere and pushed surface temperatures even higher. This also has an effect on the carbon cycle. Usually, CO_2 in the air is transformed via the Urey reaction into limestone. The CO_2 remains trapped there as the limestone is subducted by plate tectonics into magma. During a volcanic eruption, the CO_2 trapped within that magma can be released back into the atmosphere. So the amount of plate and volcanic activity on a planet can affect its atmospheric composition. But without liquid water, the Urey reaction that turns CO_2 into limestone cannot occur, trapping CO_2 in the atmosphere and providing further warming in a runaway greenhouse.
Mars, on the other hand, has a very thin CO_2 atmosphere. It has a lot of diurnal temperature variations, but overall has very cold surface temperatures. Mars, too, may have started out like Earth. However, its atmosphere has since vanished, leaving it in its cold state.

4 Other Solar System Objects

Also in our solar system we have gas giants, like Jupiter and Saturn, which are governed entirely by fluid dynamics, thermal structure, and atmospheric chemistry. They have no solid surface to absorb or hold in heat, so heat from the interior easily escapes to space.
The ice giants, Uranus and Neptune, receive very little solar energy. They are covered with a slushy water-ammonia mixture.
Other satellites of interest are Titan, Saturn's moon, and Europa, a moon of Jupiter. Titan has a thick nitrogen atmosphere, with methane rain. Europa's surface is covered with rocks of ice, but with a liquid water ocean underneath.
These sorts of climates, although very different from our own, are worth considering.

5 Extrasolar Planets

We can also look at the planets around other stars. We classify these other stars by their luminosity, which increases over time, and their temperature, which determines the type of radiation they emit. We mostly look at main sequence stars, because they tend to be more stable than stars leaving the main sequence. Of course, when we look for planets around stars, our sample is skewed towards the planets that are easier to detect.

When investigating the climate of these planets, we need to consider the amount of energy the planet gets from its star, as well as the planet's atmosphere, composition, and size. We also have to look at the orbits: eccentric orbits can yield novel seasonal cycles, while tidally locked planets provide a new challenge as their climates depend on how energy transfers from the day side to the night side.

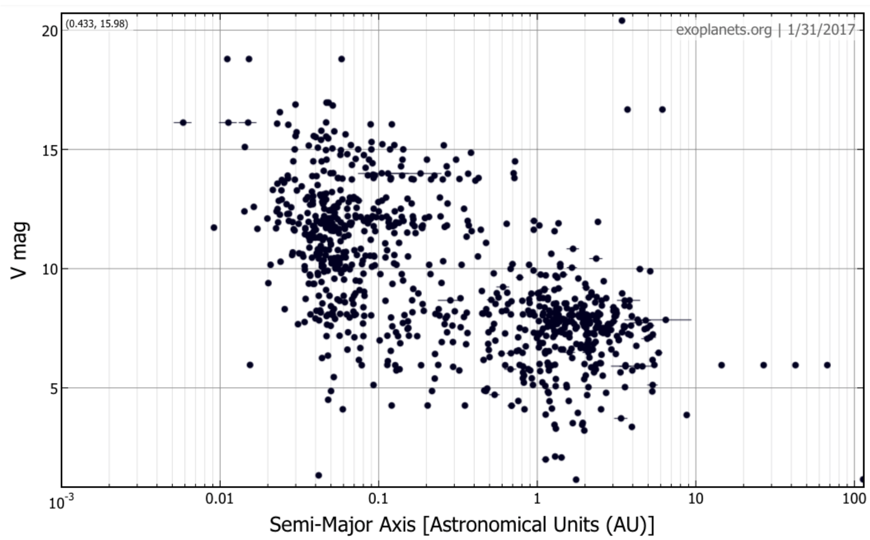

Figure 2: A scatter plot of star magnitude in the visual spectrum versus orbital distance taken from current exoplanet data. These are both properties that one might like to explore in relation to a planet's climate. The magnitude data gives us information about how bright the star is, and the semi-major axis gives us an idea of how much of its light will reach the planet. There is a huge range in the properties of exoplanets. This data can be accessed at: exoplanets.org. From the plots section of the website one can compare and plot any of a long list of different properties of the cataloged exoplanets and their stars.

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6 Climate Proxies

Climate Proxies give us information about what the climate looked like at some time in the past. These are often things preserved in the geologic record.

Some climate proxies are:

- Particular rocks or rock formations that can only be formed in certain ways, such as fossilized river networks that tell us about the presence of liquid water, or rocks that reveal the presence of glaciers.

- Biological fossils preserved in rocks, (during eras where there was sufficient biological diversity) if we can infer the biological requirements of those organisms.

- Deep ocean marine deposits, when they end up on land, can tell us about the climate based on the sediments found in it.

- Chemical and biological indicators, such as the temperature dependent compositional ratios of corals or certain microorganisms.

6.1 Isotopic Proxies

We can analyze isotopic ratios to tell us about the age of materials, or the conditions in which they formed. When certain unstable isotopes decay, we can measure the ratio of the parent to the daughter, and, based on the length of the half-life of that isotope, use this to determine the age of a material. This is how carbon-dating works.

Looking at ratios of stable isotopes can tell us about specific processes that might sort the isotopes unevenly. For example, lighter isotopes tend to move more quickly and evaporate faster. We can use this to determine what portion of an atmosphere has escaped into space. Different isotopes also can go through chemical reactions at different rates depending on the temperature, so we can use them as a paleothermometer. Photosynthesis is an example of a reaction that prefers lighter isotopes of carbon, so when we find a rock with a larger ratio of carbon-12 to carbon-13 we can say that it was formed from biological organisms.

6.1.1 Isotope Equilibrium Fractionation

Let us mix 2 distinct substances with different isotopic ratios. As they come in contact, they trade isotopes until they reach equilibrium. Then if r_1 is the isotope ratio in substance 1 at equilibrium, and r_2 is the isotope ratio in substance 2 at equilibrium, we say:

$$r_1 = f_{1,2} r_2 \tag{2}$$

where $f_{1,2}$ is the equilibrium fractionation temperature. Typically, it is close to 1, and it varies with the ratio of isotopic masses and with temperature, meaning we can use it as a paleothermometer.

6.1.2 Isotopic Compostion

We describe the ratio of isotopes r_A as the number of atoms of isotope A over the number of atoms of the dominant isotope. We then also define r_{AS} as the isotopic ratio for a standard reference sample, because differences in isotopic ratios are easier to measure than absolute values. Then the isotopic composition of sample A is:

$$\delta_A = \frac{r_A - r_{AS}}{r_{AS}} \tag{3}$$

The δ value is usually expressed as a parts-per-thousand value, denoted . If δ_A is negative, our sample is depleted in isotope A relative to the reference.

Given these equations, we can start to look at some relevant examples. We examine the isotopic ratios of carbon-13, because it tells us about photosynthesis on Earth. Carbon on Earth is outgassed from volcanos and magma, and this process is balanced by the burial of carbon, either in the form of inorganic carbonates or organic carbon. The organic, photosynthetically produced produced carbon has $\delta^{13}C$ values 25lower than that of the reservoir.

We add up the mass balance of all the carbon. If f_{org} is the fraction of carbon which is buried in the form of organic carbon, δ_o is the $\delta^{13}C$ of carbon outgassed from the Earth's interior, δ_{org} is the $\delta^{13}C$ of the organic carbon buried and δ_{carb} is the $\delta^{13}C$ of the non-organic carbon buried, we have:

$$\delta_o = f_{org}\delta_{org} + (1 - f_{org})\delta_{carb} \tag{4}$$

If we have data on the isotopic ratios for both types of sediments, we can infer f_{org} to learn about the amount of organic carbon on Earth. This tells us about the state of the carbon cycle, provided it is in equilibrium. If it isn't, doing a detailed accounting of the flows of carbon can provide us information about what might have disrupted the cycle.
There are many other ways that we can use isotopic information to learn about Earth's climate. For example, isotopes of water are used as a proxy to find out how much ice was present on Earth. This is because lighter isotopes of water evaporate more easily than heavier isotopes. The water vapor in the air is therefore isotopically lighter than the ocean water. When this vapor condenses, precipitates over land, and forms glaciers, the light water is stored, making the ocean water noticably heavier. If we measure the isotopic composition of the ocean at a certain time, we can calculate how much ice there was on Earth at that time. But how do we acquire such information about past oceans?

6.1.3 Forams

Foraminifera, or forams for short, are single-celled shelly amoebas. Their $CaCO_3$ shells have different shapes depending on the isotopic composition of the water they live in. When they are fossilized, we can examine their shells to find out about past states of the ocean. Measureable fossils go back about 70 million years. The oxygen in the shells is also isotopically fractionated with respect to the ocean, and this factor changes with temperature. This means that forams can also be used as paleothermometers, however, it is hard to disentangle the temperature effects from ice-volume effects on isotopic composition. Despite this complication, forams still provide valuable information about past climates due to these and other chemical signatures.

7 Earth's Climate History

During the Proterozeric, which spans from 2.5-543 million years ago, Earth's climate and atmosphere was adjusting to the presence of microbes. Photosynthesis allowed oxygen to accumulate in the atmosphere, and since oxygen can be very reactive, this also allowed CO_2 and H_2O become dominant in the atmosphere, instead of CH_4. However, many bacteria do a reaction that takes O_2 out of the atmosphere and turns it in to CO_2. So in order for oxygen to accumulate into the atmosphere, the organic carbons formed from photosynthesis must be buried before it can be re-oxidized. In analyzing the Proterozoic, we see many swings in the $\delta^{13}C$ of carbonates. We see evidence of major glaciations as well. The causes of these events have to do with the carbon cycle, the oxygen cycle, the sulfur cycle, and other aspects of biogeochemistry. We also see evidence of major glaciation events, including 'snowball' states where ice sheets reached to the tropics, which only ended with a buildup of CO_2 in the atmosphere to cause warming.
Moving into the Phanerozoic, our current eon, we begin to see multicellular life diversifying. During this period, we see a series of swings from warm climates, where the oceans were ice free and the planet had a relatively small temperature gradient from tropics to poles, to icy climates, where there were large amounts of ocean ice and ice sheets. These ice ages could have been due to fluctuations in greenhouse gases in the atmosphere over long time scales. We also see evidence of a massive, abrupt, and transient warming event around 55 million years ago. This event, called the PETM for Paleocene/Endocene Thermal Maximum, was possibly caused by a release of lots of carbon into the atmosphere. We can see from the fossil record a mass extinction of forams, probably due to warming, ocean acidification, and oxygen depletion. From the hothouse/icehouse swings in the Phanerozoic, we can see an interesting example of how geography, such as the formation of supercontinents, affects climate. It is easier for glaciers to form if there are landmasses near the poles, and the presence of ice-free landmasses allow the silicate weathering of rocks that helps govern the CO_2 levels of the atmosphere. The shape of the land and the ocean also affects the ocean's ability to

transport heat. When there are large supercontinents, less of the land is near the moderating effects of the ocean, allowing larger seasonal temperature swings.
During the past 5 million years, the Earth has experienced a fairly regular cycle of ige ages and interglacial periods. The climate tends to cool over long periods and then warm up relatively quickly. Some of this periodicity can be attributed to Milankovic cycles, which are periodic changes in Earth's orbit and axis tilt that affect the amount of light we get from the sun. Starting about 1 million years ago, we can obtain data about Earth's past by examining ice cores. By drilling deep into ice sheets and analyzing the isotropic ratio of the water as well as the CO_2 trapped in ice bubbles, we can confirm warming and cooling glacial cycles, and we can see that CO_2 levels rise and fall with warming and cooling.
This brings us to the Holocene, our current period, starting around 12,000 years ago. For the most part, the Holocene climate has been very stable, with more subtle variations than we have seen in previous times. As we get closer to present time, however, we need to examine the impact humans have had on the climate.

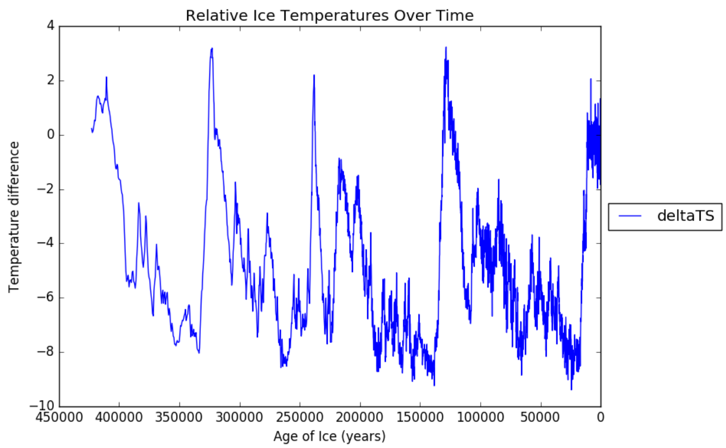

Figure 3: A graph of relative temperatures, as measured by ice core data, over time.

8 Global Warming

We have seen that the carbon cycle has a large influence on climate. Humans have done a lot to change the carbon cycle, most prominently post-industrialization by burning hydrocarbons and releasing greenhouse gases like CO_2 into the atmosphere. In 2012, an estimated $3.56 \cdot 10^{13}$ kg of carbon were released into the atmosphere, as compared to the background volcanic outgassing of around 10^{11} kg. The human impact on the carbon cycle dominates natural impacts. CO_2 can remain in the atmosphere for centuries or millennia, and this has an effect on Earth's temperature. The exact temperature change we can expect is hard to predict, because it depends on how much CO_2 is absorbed into the oceans, how much feedback there is in the atmosphere from evaporating water vapor, how air pollution affects the clouds covering earth, and how much changes in snow, ice, glaciers, vegetation, and ocean currents affect the temperature.

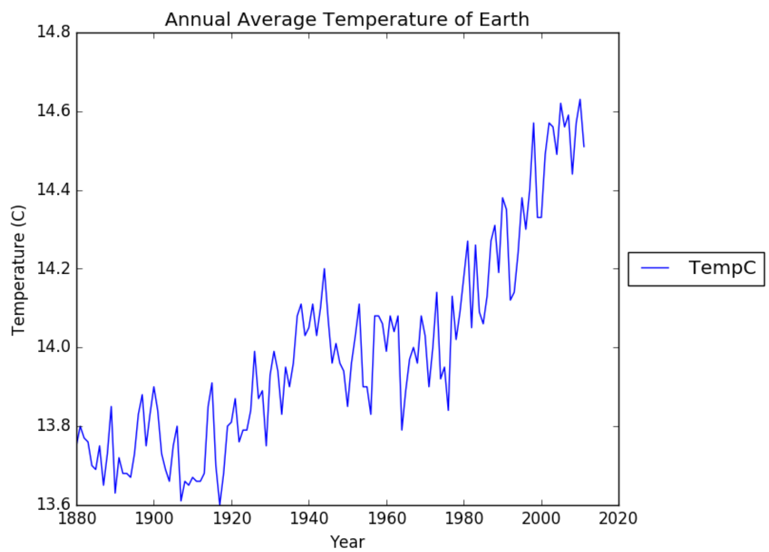

Figure 4: This graph shows average temperatures on Earth from 1880 up until 2011. Despite fluctuations, there is a clear upward trend in the data.