Exploring the World of Science

2026 Remote Sensing C
Content Introduction

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Contents

1 Introduction

- This document was partially created based on draft rules, and so may not be 100% up to date. Please refer to the official Science Olympiad 2026 Rules Manuals as well as Rules Corrections and Clarifications (all available via the national website `https://soinc.org`) for the finalized event topics and rules of competition.
- Every year there are several hundred Science Olympiad tournaments hosted nationwide, which means several hundred Remote Sensing C event supervisors and several hundred interpretations of the content covered in the event rules. While this document aims to provide an overview of the fundamentals of the event topics, remember that the rules manual (including rules corrections and clarifications) is the official source, and the topics covered here are ultimately just a result of the author's interpretation of the rules.
- Like any Science Olympiad event, the topics covered in Remote Sensing C can become quite deep and expansive, especially at higher levels of competition. This document tries to focus on the most fundamental parts of the event, with the goal that after the reader familiarizes themselves with these topics they will have a strong foundation and be able to independently deepen their knowledge.
- Remote Sensing rules vary significantly between Division B vs. Division C, so while this document may occasionally be useful for Division B, it's not the focus and won't be covering the same topics.

2 Instrumentation and Physics

The foundation of remote sensing begins with the physical processes that we use to make observations - primarily different wavelengths of light, and how they interact with objects that we want to observe - and the instrumentation that we use to make observations. This section covers some of the fundamental concepts of this topic.

2.1 Interactions of Light and Instruments

Key to many methods of remote sensing is the ways in which light interacts with objects and substances of interest, and the information conveyed by these interactions.

Passive sensing relies on radiation from either the Sun (often called "shortwave radiation"), or the Earth (longwave - the origin of these names is explained in the later subsection "Blackbody Radiation"). Various objects in the Earth's atmosphere and surface reflect shortwave radiation, to greater or lesser extent. For example, the ocean reflects a very small proportion of shortwave radiation (less than 10%), whereas ocean ice reflects a much greater proportion, often more than half. A satellite that is measuring shortwave radiation - or more often, using a radiometer that is sensitive to some narrow band of wavelengths that fits within common shortwave wavelengths - can use this reflection proportion property to distinguish between ocean water and ice.

The process of reflection by molecules in the atmosphere can also reduce ("attenuate") the amount of light that makes it to the other side. Suppose our shortwave-observing satellite was looking at a perfectly reflective surface on the Earth. The satellite would measure less shortwave radiation compared to the amount that entered the top of the atmosphere as sunlight - some of it would be attenuated on the way down and up. Mathematically, attenuation is represented as exponential absorption, similar to radioactive decay - a certain proportion of the radiation will be attenuated across a unit distance. For example, suppose we pass a beam of light with intensity of 1000 W/m^2 through the atmosphere for 100 km. If the atmosphere attenuates 1% of the light for every kilometer of travel (i.e. 0.01 attenuation per kilometer), across 100 km this original beam will attenuate to 1000 $W/m^2 \cdot (1 - 0.01)^{100} \approx 366\ W/m^2$

2.2 Spectroscopy

Light is classified according to the electromagnetic spectrum, as shown in this image.

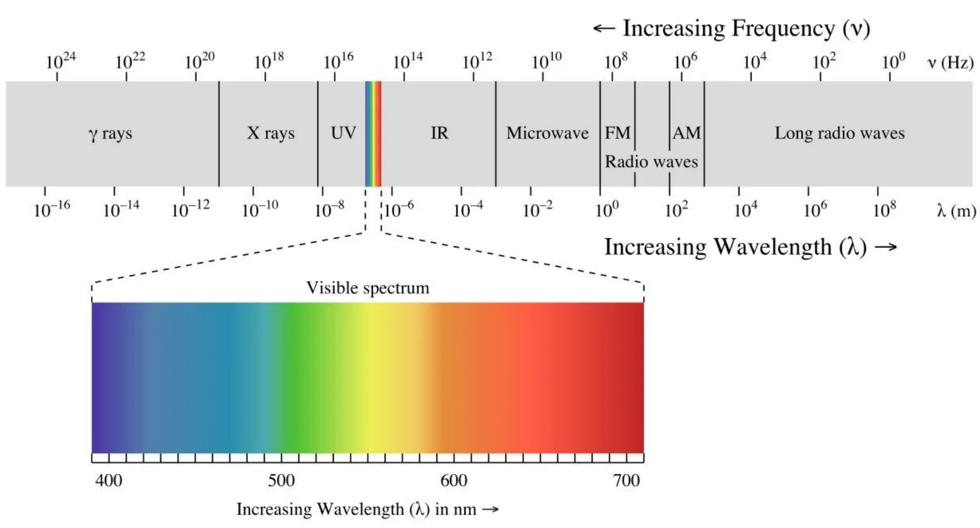

Image 1.1 - The electromagnetic spectrum. Many properties of light are covariant across the spectrum. For example, shorter wavelengths of light also carry more energy, and tend to be emitted more by higher-temperature objects.

The electromagnetic spectrum is particularly relevant to remote sensing because different molecules tend to interact with specific wavelengths of light, absorbing or emitting in a very narrow band of wavelengths. Spectroscopy is the study of these interactions, and the spectral lines produced by interactions can tell us a lot about molecular compositions. For example, consider the following absorption spectrum - in this case, from a sensor that is observing light as it passes through the atmosphere.

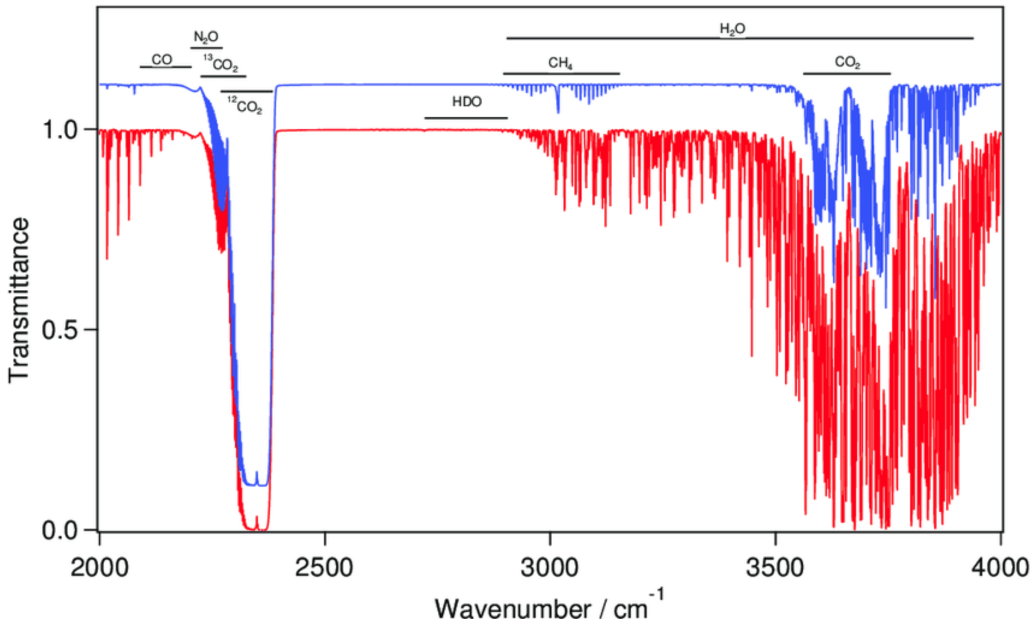

Image 1.2 - An absorption spectrum of the atmosphere. The x-axis shows wavenumber, the inverse of wavelength. Red is regular air, and blue is dry air.

Notice how transmittance drops substantially in narrow bands, where different molecules absorb radiation of specific wavelengths. For example, around wavenumber of 2350 cm^{-1} (wavelength of approximately 4.26 microns), carbon dioxide absorbs almost all of the radiation. This means that if a satellite wants to detect the presence of carbon dioxide, having a sensor that is tuned to that specific range can be a good option. For example, the AIRS instrument on the Aqua satellite (which is one of the satellites listed in the event rules this year) takes measurements near 4.26 microns as part of its carbon dioxide observations.

2.3 Blackbody Radiation

Radiant (light-emitting) objects are often modeled as a blackbody - a uniform object that emits light due to its own temperature, but does not reflect any incoming light. Many physical objects are very similar to theoretical blackbodies - for example, the Sun - and the blackbody model provides a starting point to attach the extra complexity of reflection, multi-layered surfaces, etc. This subsection covers the fundamentals of working with blackbodies - that extra complexity will be discussed in the "Energy Balance Modeling" section later in this document.

The convenience of the blackbody model is that the light emitted by a blackbody depends only on one property - its temperature. A blackbody emits a radiant flux - energy output per unit surface area, in the form of light - according to the Stefan-Boltzmann Law: $Q = \sigma T^4$ where Q is the radiant flux (Watts per square meter), σ is the Stefan-Boltzmann constant $\sim 5.67e-8\ W{\cdot}m^{-2}{\cdot}K^{-4}$ (using scientific notation for convenience), and T is the temperature (Kelvin). For example, the surface of the Sun has a temperature of approximately 5800 Kelvin, which means that if we model it as a blackbody, its radiant flux would be $\sim 6.44e7\ W/m^2$. Thanks to the power-of-4 in the formula, energy output scales very quickly with increasing temperature - twice the temperature means 16 times the energy output.

The distribution of the light emitted by a blackbody forms a curve across the electromagnetic spectrum. The peak emission wavelength is determined by Wien's Law: $\lambda = b/T$ where T is temperature in Kelvin, b is Wien's displacement constant $\sim$2898 $\mu m{\cdot}K$, and λ is the peak wavelength (measured in microns, same as the component unit of the constant value).

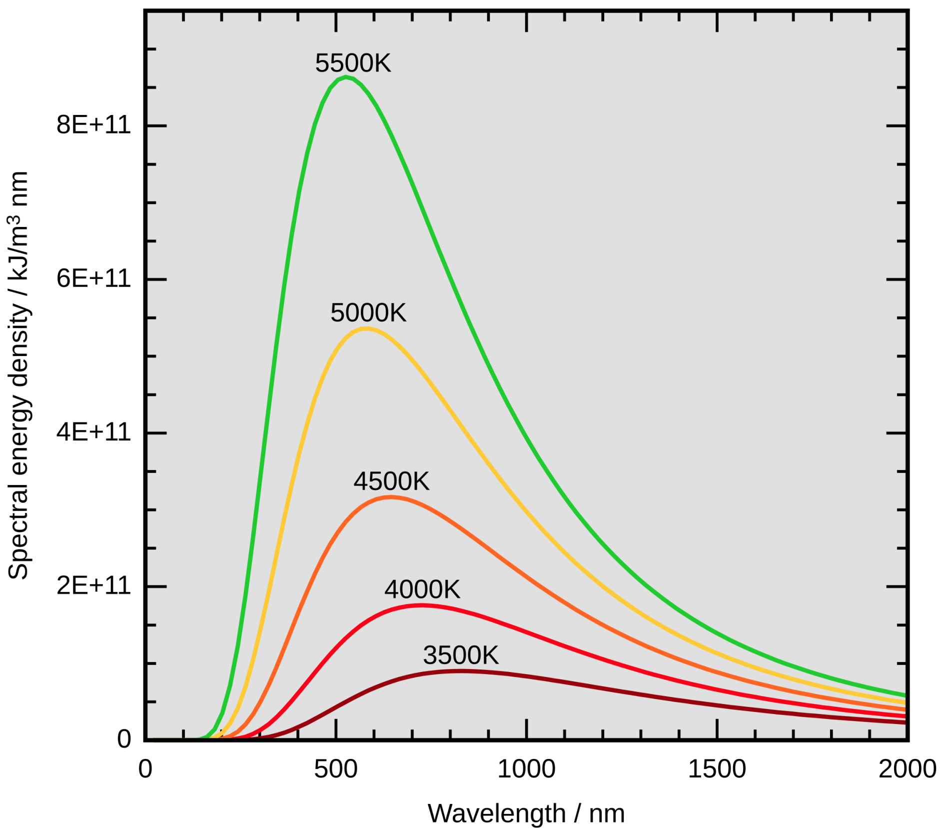

Image 1.3 - Emission curves for various temperatures. Notice how hotter objects shift their peaks toward shorter wavelengths, but still emit more light across the entire spectrum.

As shown in Image 1.3, many objects emit only a small proportion of their light in the visible range. For example, a room-temperature object (e.g. the Earth as a whole) can still be modeled as a blackbody, but it emits almost no visible light - most of its emissions are in the infrared portion of the spectrum.

Blackbody radiation is the reason why solar radiation is called shortwave and Earth's radiation is called longwave. Per Wien's law, the temperature difference means that the Sun's radiation tends to be at shorter wavelengths, while the Earth's radiation tends to be at longer wavelengths, as shown in the image below.

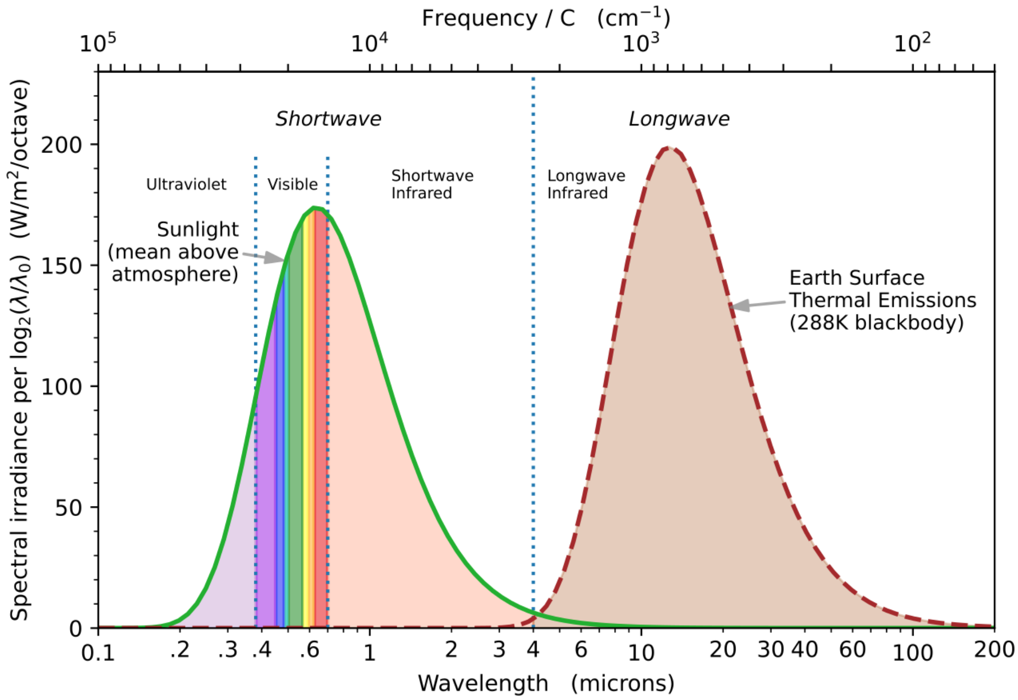

Image 1.4 - Comparison of shortwave radiation at the top of the atmosphere vs. longwave radiation emitted by Earth.

Note that this image shows shortwave radiation at the top of the atmosphere. The radiant emission at the surface of the Sun is extremely high compared to the Earth's emissions, but the Sun is very far away from the Earth, so the emitted light spreads out, and the radiative flux near the Earth is only $\sim$0.002% of what is emitted at the surface of the Sun.

2.4 Orbital Properties

The aspects of remote sensing that are relevant to this year's rules are satellite-based, so the mechanics of satellite orbits are necessary to understand the benefits and drawbacks of various orbital patterns. One of the lowest-order effects at play in orbital mechanics is Kepler's Third Law, which describes a rather strict relationship between orbital altitude and orbital period. We can express it mathematically as $T^2 = 4\pi^2 G^{-1} M^{-1} r^3$ where T is the orbital period, r is the distance between masses, G is the gravitational constant $6.67e-11\ N{\cdot}m^2/kg^2$, and M is the sum of the masses of the two objects (which for our purposes, for satellites is almost identical to the Earth's mass). The key insight is that for any particular altitude, there is only one possible orbital period, and for any orbital period, there is only one possible altitude. If an orbiting satellite speeds up or slows down, it will necessarily decrease or increase the altitude of its orbit. This law can be used to compute the altitude at which a satellite must orbit in order to have a particular orbital period - for example, all geosynchronous satellites orbit at the same altitude.

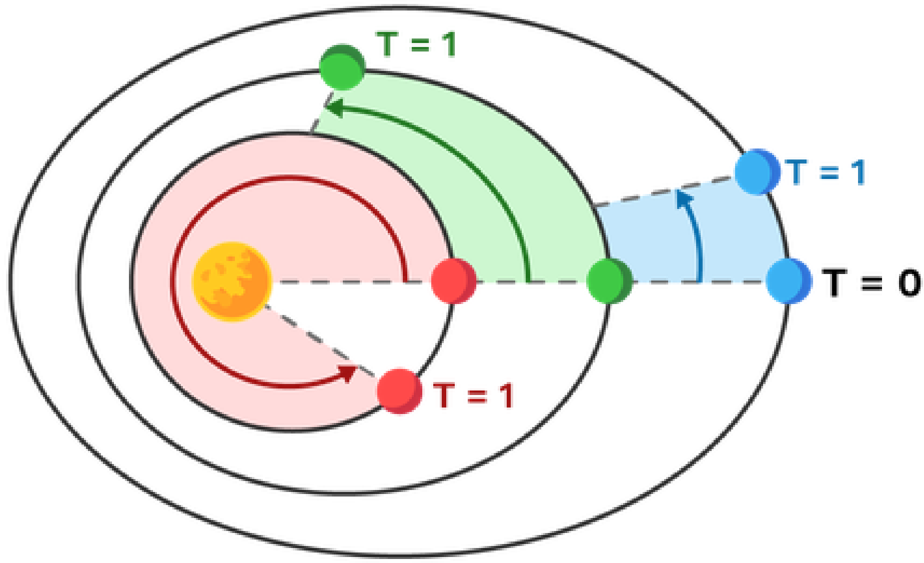

Image 1.5 - Kepler's Third Law, illustrating that greater orbital distance corresponds to longer orbital period, at a ratio of $T^2 \propto r^3$.

Satellites orbit at various altitudes for different purposes. The core divisions are Low Earth Orbit, Medium Earth Orbit, and High Earth Orbit, where the latter two in particular are divided by the geosynchronous altitude (where satellites have an orbital period of one day - per Kepler's Third Law, everything in Low and Medium Earth Orbit has a shorter period, and everything in High Earth Orbit has a longer period). There are a wide variety of satellite types placed in these various orbital regions, but in general satellites are placed in the more expensive-to-reach higher orbits either due to the longer periods (for example, geostationary satellites like GOES that want to observe the same region indefinitely), or, less relevantly to this event, because they are able to make observations that would not be possible in a lower orbit, e.g. if they are blocked by the Earth's magnetic field.

In addition to the effects discussed so far, there are many other properties of orbits that may influence the orbital pattern choices made for a satellite. For example, many of the satellites listed in the rules follow Sun-synchronous orbits - where they experience the same solar time, and therefore same angles of shortwave radiation, each time they revisit a particular location - which depends on higher-order effects such as the precession of the orbit.

2.5 Imaging Resolution

The purpose of satellite imaging is to collect high-quality data, so we must be able to understand the tradeoffs in resolution - spatial, temporal, etc. - that satellites must make, and how these tradeoffs relate to an observation mission's goals.

When considering spatial resolution, the key point to keep in mind is that ultimately, remote sensing data is a matter of having one or more sensors pointed at targets that measure incoming light in a particular spectral range. Consider, for example, an ordinary camera. When you take a picture, the camera acts like an array of sensors, each corresponding to a single pixel and color channel, which measure the light that hits them. The farther you are from the target, the lower your spatial resolution, because each sensor covers a wider area - the instantaneous field of view (IFOV) of the sensor (usually a solid angle measurement) is spread out over a larger area (linearly with distance, as an inverse-square law for area). If you try to view two things in the same IFOV, you won't be able to tell them apart, because the single sensor corresponding to that IFOV can only measure the total of them.

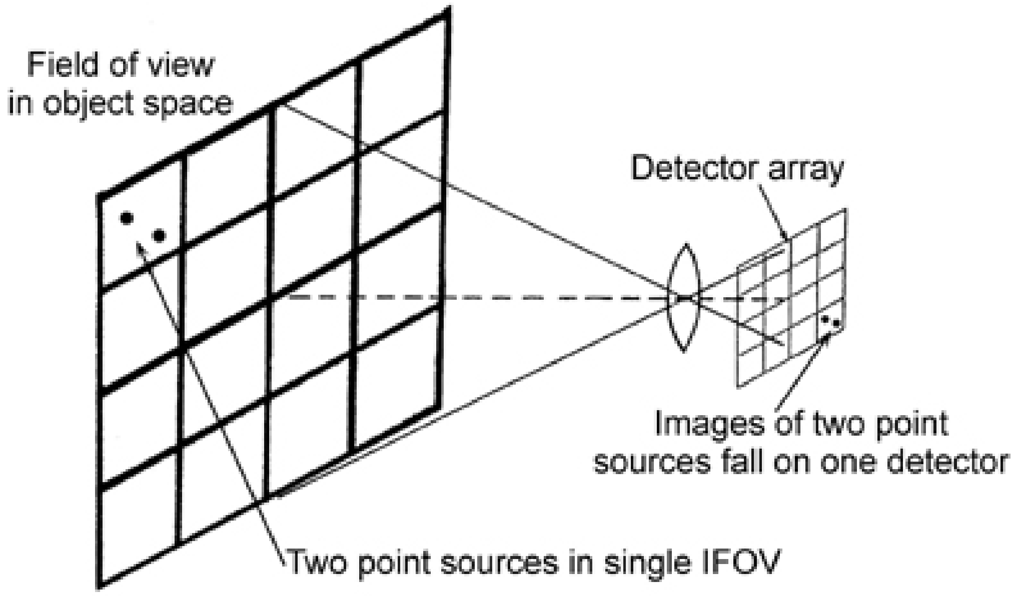

Image 1.6 - Instantaneous Field of View and its correspondence to imaging.

As such, spatial resolution of satellite sensors - the ground sample distance, i.e. how close together things can be before they are not reliably resolved - depends on both the IFOV of the sensors themselves, as well as the distance of the satellite from its targets on Earth.

The most well-known form of resolution is spatial, but other aspects are just as important. Consider temporal resolution - the time between subsequent imaging of a location, a common constraint on satellite missions. For example, a satellite like OCO-2, which is intended to track carbon dioxide in the atmosphere over time. It has a revisit period of 16 days, which balances being frequent enough to measure seasonal variations, with the engineering limitations of its orbital pattern and spatial resolution, as a satellite with OCO-2's configuration (relatively narrow swath width, frequent imaging due to low usability of raw data, operating in a pushbroom mode) must make many orbits around the Earth at slightly different offsets before it can revisit a location. In contrast, a satellite like GOES-16 prioritizes high temporal resolution - it is placed in a geostationary orbit, so it trades off only being able to cover a portion of the Earth's surface, but provies high spatial resolution and a "revisit" period of only a few minutes. Each mission made different choices when considering its goals and the tradeoffs between various types of resolution.

3 Image Interpretation

Remote sensing as a concept is fundamentally about data collection and interpretation, so examining imaging data is key to this event. The rules lists several groupings of topics alongside many specific satellites and instruments for the event - here, we'll approach the idea of image interpretation from a general perspective, and show a couple of examples to illustrate how to apply your understanding of the event topics to image interpretation.

3.1 Principles

The core of image interpretation as a competitor in Remote Sensing is a single question - why? The spectral range(s) of an image have a reason for their use - some scientist or engineer made the choice to put sensors for those specific wavelengths because they wanted to study some useful information. Someone (maybe a test writer, maybe not) made an image from a specific set of data because they wanted to show something interesting in that data. To succeed with image interpretation, examine the patterns of the image, and use the data and your knowledge of remote sensing principles to see through the eyes of the event supervisor.

3.2 Example - Sea Surface Temperature

Consider the following image produced by Jason-3

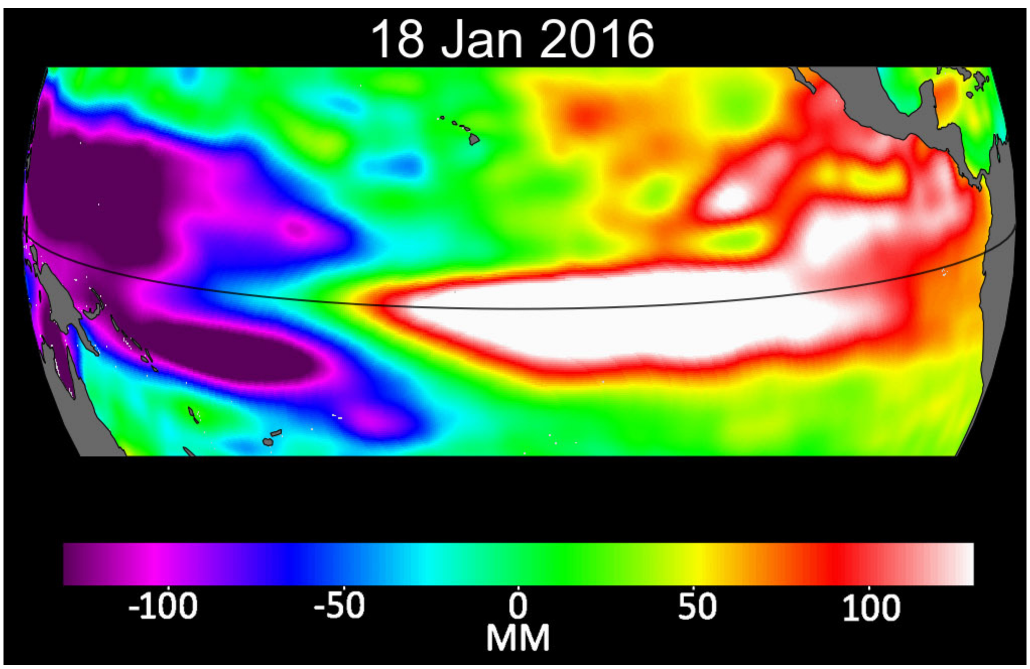

Image 2.1

Let's consider some aspects of this image:

- The scale on the bottom of this image is measured in millimeters. Notice that this scale is centered around zero - it is a measurement of anomaly, or the difference from some normal baseline.
- Jason-3 uses radar altimetry to measure anomalies in sea surface height. The most common purpose of measuring sea surface height is as a proxy for sea surface temperature (because warmer water exerts more pressure, and so rises a bit higher on a regional scale).

- We can observe that there is a clear lower-than-baseline height (cooler-than-usual waters) in the left side of this image, and vice versa on the right side.
- Considering the previous observation, notice that the region that this image shows is the tropical Pacific Ocean, which is a key driver of the El Niño-Southern Oscillation. In particular, the El Niño phase involves relative warming in the East Pacific.

If you saw this image on a test and were able to figure out this information, you may already have a good idea of what the questions relating to the image are likely to be about.

3.3 Example - Atmospheric Carbon Dioxide

Consider the following image produced by OCO-2

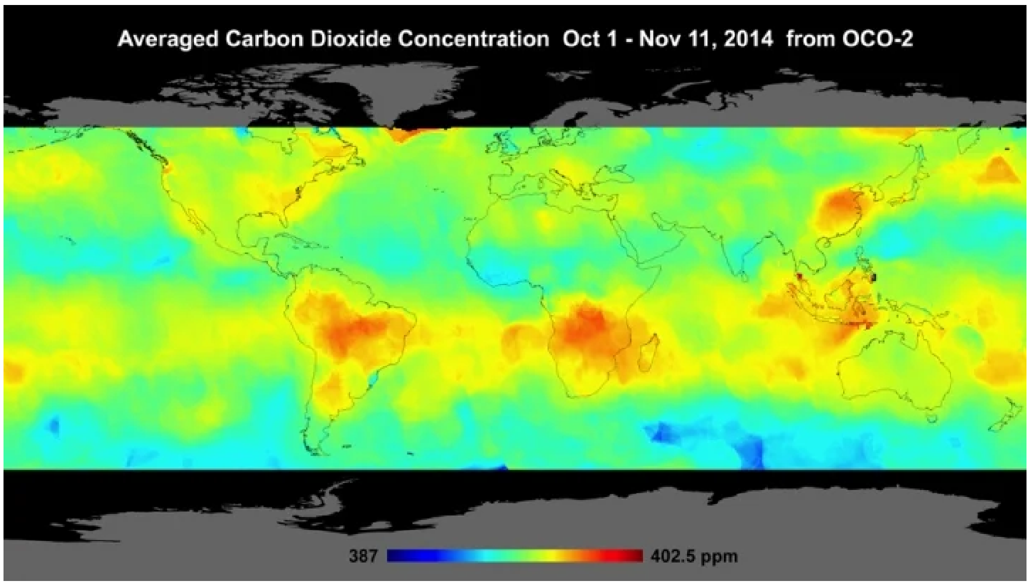

Image 2.2

Let's consider some aspects of this image:

- Carbon dioxide in this image is being measured in ppm - parts per million, which is the most common unit for CO2 measurements. (Sidenote: there are nuances to this unit that may sometimes be relevant, such as that it is a molar measurement, and that it disregards atmospheric water vapor in the denominator)
- Notice that there is no data in the polar regions. Coverage is limited or absent for OCO-2 when the Sun is at narrower angles. While OCO-2's Sun-synchronous orbit allows it to maintain the same local time, this can't compensate for the solar path angle differences at high latitudes.
- This image provides a date range when this data was collected. Seasonal changes correlate strongly with local carbon dioxide levels, for various reasons. For example, due to summertime vegetation growth, carbon dioxide levels in the Northern hemisphere are at an annual low point in this image. Since the vast majority of land area and vegetation are in the North, this means that global average carbon dioxide levels are also near an annual low point at this time.
- Localized anthropogenic emissions that have not yet dispersed more broadly are also visible around many regions. The precise drivers of these sources vary from place to place.

Applying preexisting event knowledge to new scenarios is at the core of interpretation. The primary barrier in many cases is to avoid being intimidated, so that you can confidently break down the data into contributing aspects, and rebuild the whole to produce a cohesive picture.

4 Climate Processes

4.1 Atmospheric Gases

While many simple models treat the atmosphere as a uniform entity, in reality it contains many different trace substances in varying quantities, each of which have different effects on overall atmospheric and surface conditions.

Greenhouse gases are particularly relevant to the Earth's average temperatures, as well as having some additional local effects. Greenhouse gases in the atmosphere trap heat, raising the Earth's temperature the more there are (the mechanism is described further in the later subsection "Albedo, Absorption, Reemission, and Radiative Forcing"). Of course, not all greenhouse gases are the same - they vary substantially in the particular wavelengths they absorb, the strength of their greenhouse effect, and the length of time they stay in the atmosphere before being converted to other molecules (whether atmospheric or in the surface). For example, methane is a much stronger greenhouse gas than carbon dioxide, but does not last as long in the atmosphere (though it also tends to decompose into carbon dioxide itself). Many greenhouse gases also have substantially seasonality in their local concentrations, complicating measurements further.

In addition to greenhouse gases, the atmosphere also contains various other trace gases that are relevant to climate processes. For example, sulfur dioxide, a common emission from volcanoes and anthropogenic sources, forms aerosols in the atmosphere which reflect shortwave radiation, producing a cooling effect. Ozone in the stratosphere is a significant factor in the level of UV radiation that reaches the Earth's surface, and its levels can be affected by several different trace gases, including some listed in the rules.

4.2 El Niño-Southern Oscillation

In addition to seasonal variation and long-term climate change, the climate also varies on less regular scales. The El Niño-Southern Oscillation is one of the most relevant such climate variability phenomena - an irregular shift in climactic conditions caused by changes in the Pacific Ocean.

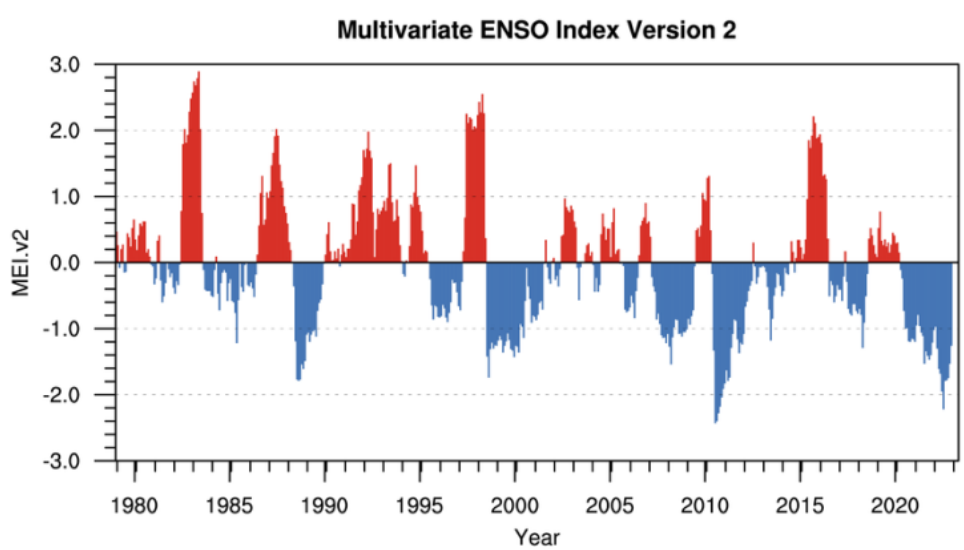

Image 3.1 - Historical El Niño-Southern Oscillation fluctuation.

To understand the ENSO, we'll start by looking at the typical state of the tropical regions of the Pacific Ocean. Ocean currents and atmospheric circulation at the surface in this region is zonal (in the direction of latitude lienes), specifically moving East to West, with cooler waters brought from deeper in the ocean in the East, and warmer waters after surface exposure in the West. The rainfall and wind patterns brought about by this circulation is a core element of various aspects of climate in this region, with impacts on adjacent areas as well.

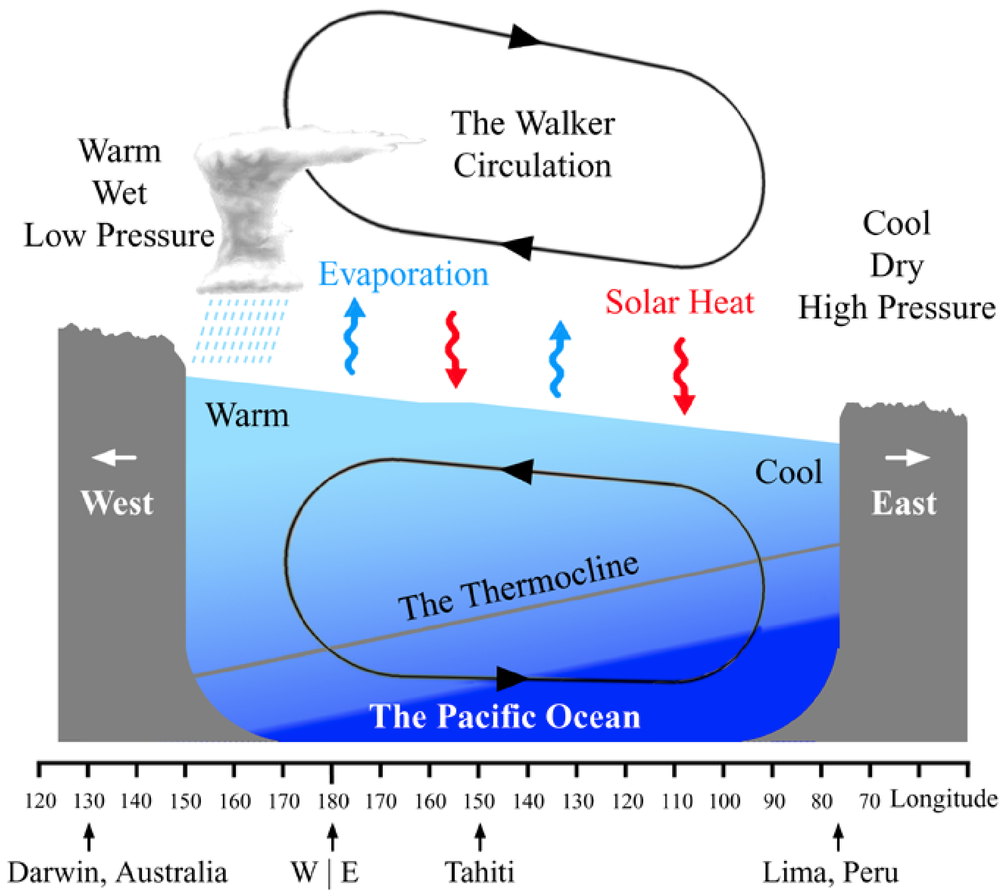

Image 3.2 - Pacific Ocean currents and atmospheric circulation.

Taking this view of things, El Nino occurs when this circulation pattern weakens, and La Nina occurs when it strengthens. These changes affect the ocean and air temperatures, wind patterns, precipitation, and many higher-order phenomena that result from these shifts (for example, changes in the productivity of offshore fishing).

4.3 Carbon and Nitrogen Cycles

We have previously discussed how atmospheric gases can affect climate change, but now we will examine the key processes that describe the atmospheric concentrations of many of those gases - the carbon and nitrogen cycles - and the factors that influence the movement of carbon and nitrogen between the nodes of these cycles.

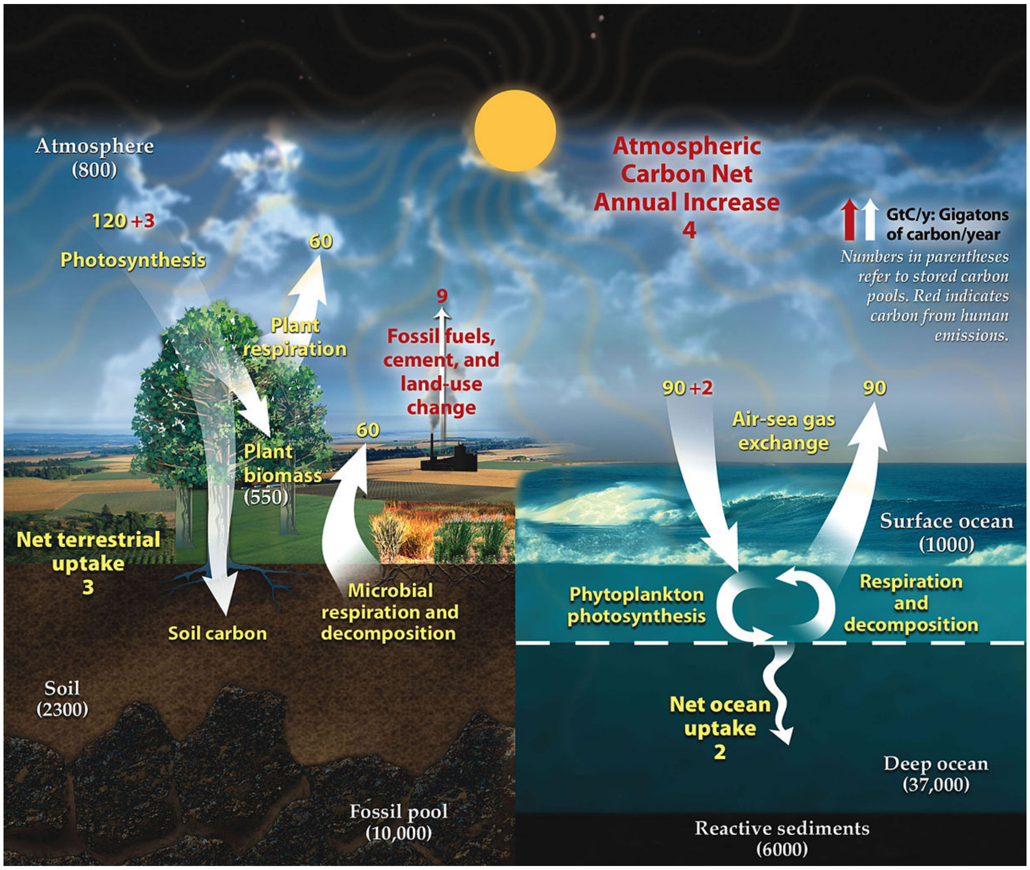

Image 3.3 - Carbon cycle.

The carbon cycle describes the movement of carbon between the Earth's surface, atmosphere, and living organisms. The key point of relevance to climate processes is that the predominant atmospheric form of carbon is carbon dioxide, a major greenhouse gas. Using the image above, notice how several natural processes transfer carbon between different carbon sinks, such as photosynthesis converting carbon dioxide to organic molecules, and oceanic organisms converting carbon dioxide to carbonic acid. Anthropogenic sources, particularly the burning of fossil fuels and methane emissions, create changes in the strength of these processes - hence some excess carbon is absorbed by plants, or the ocean (hence "ocean acidification"), while the remainder stays in the atmosphere as carbon dioxide, driving increased temperatures.

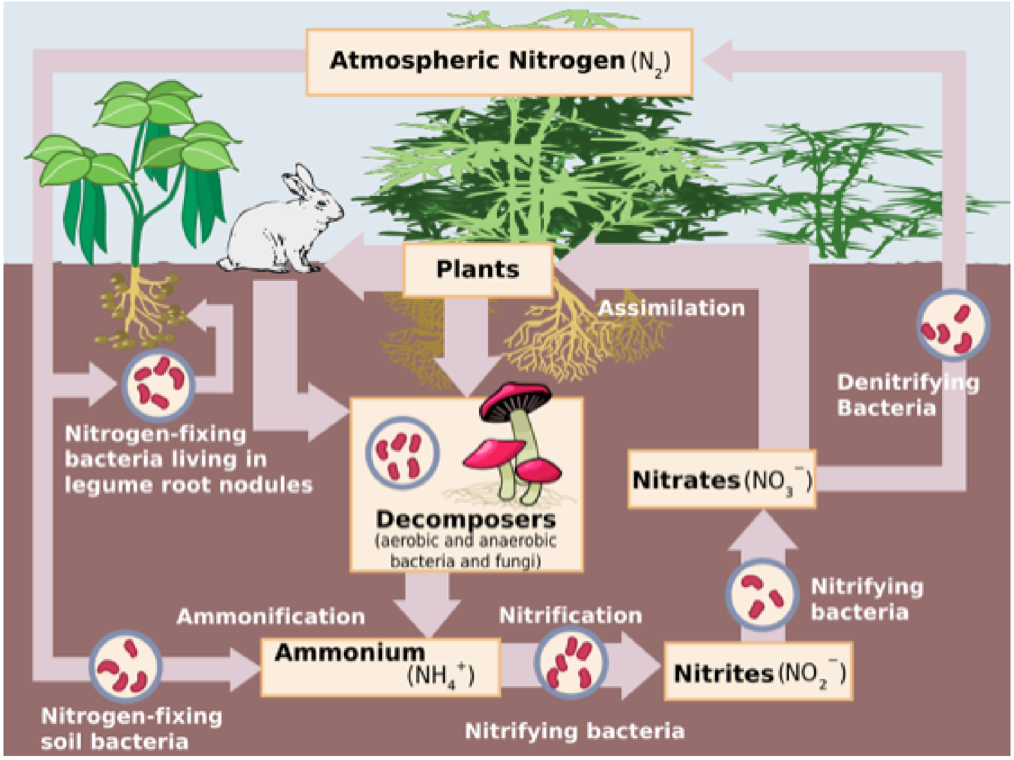

Image 3.4 - Nitrogen cycle.

The nitrogen cycle is the corresponding model of the movement of nitrogen. Most atmospheric nitrogen is molecular, i.e. N_2, but some of it is nitrous oxide (N_2O), which has relevance both as a greenhouse gas and through its interactions in the stratosphere where it breaks down ozone, reducing the effectiveness of the ozone layer in blocking harmful UV radiation from reaching the Earth's surface. Conversion of nitrogen between different forms in the Earth's soil by microorganisms and plants is a key aspect of plant growth patterns, and nitrogen fertilizer used in agriculture is a major source of human contributions to nitrous oxide in the atmosphere.

Of course, there are many other factors beyond those listed here that are relevant to the workings of these cycles. You might see questions about how changes in the processes moving carbon and nitrogen between these sinks affect the strength of flows among them.

4.4 Effects of Climate Change

The primary sense in which we talk about "climate change" tends to be increases in average global temperature, but in the Earth's complex ecosystem these temperature changes affect many other aspects of the climate. For example, global temperature increases mean that the average polar ice cover will be lower, which would tend to increase sea levels. Clouds, precipitation, and storm systems are generally driven by heat and corresponding pressure differences helping to bring water up into the atmosphere, so global temperature changes have effects on the locations, frequency, and strength of rainfall and severe weather. Plant life is heavily dependent on local temperature and precipitation conditions, which in some locations can change substantially from relatively small shifts in global temperatures, so the regions for particular types of vegetation and the viability of agriculture for particular crops may move around. Questions on this topic could relate to identifying or analyzing these shifts in data.

5 Energy Balance Modeling

The foundational model for analyzing the Earth's temperature and temperature change is a star-and-planet energy balance equilibrium model. The basic version of this model treats the Sun and Earth as uniform blackbodies, and computes the Earth's temperature based on the amount of energy emitted by the Sun, and the sizes and distance between the two bodies.

5.1 Radiative Energy Balance

The equilibrium temperature of the Earth under this model means that the amount of energy the Earth is receiving (from shortwave radiation) is equal to the amount of energy emitted by the Earth back into space (longwave radiation).

For the Sun's temperature, we can begin by considering its surface temperature - $\sim$5800 K. Recalling the Stefan-Boltzmann formula and the example from the earlier subsection "Blackbody Radiation", we can compute the radiant flux at the Sun's surface - $\sim 6.44e7$ W/m^2, using scientific notation for convenience.

The vast distance between the Sun and the Earth means that the incoming flux is much lower. Emitted light from the Sun spreads out over time, reducing with the square of relative distance (i.e. it follows an inverse-square law), as seen in this image.

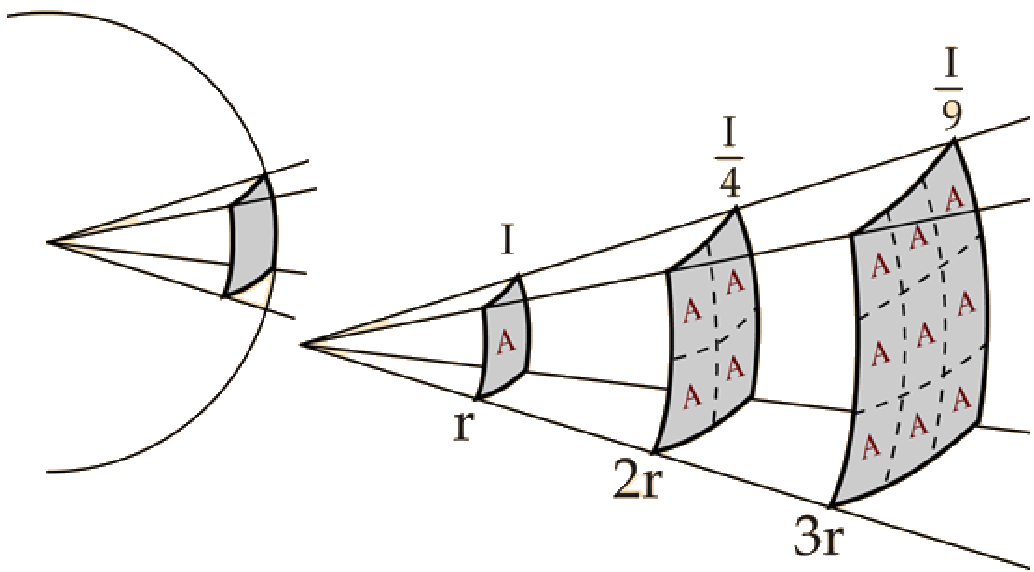

Image 4.1 - The inverse-square law that explains solar emission intensity reduction.

We can represent this mathematically as $Q_2/Q_1 = (d_1/d_2)^2$ where Q_1 is the flux at the Sun's surface, d_1 is the distance from the center of the Sun to the Sun's surface (i.e. the Sun's radius), Q_2 is the flux entering the Earth, and d_2 is the distance from the Earth to the Sun. If we rearrange this equation and plug in the real values (you can look them up yourself), we can see that the flux entering the Earth is $\sim$1284 W/m^2 under this model.

Recall the beginning of this subsection - the energy entering Earth as shortwave radiation must equal the energy leaving as emitted longwave radiation. We have computed a value for the flux entering the Earth, but we cannot use this directly, because while the Earth emits longwave radiation as (approximately) a sphere, it receives shortwave radiation as (approximately) a circle.

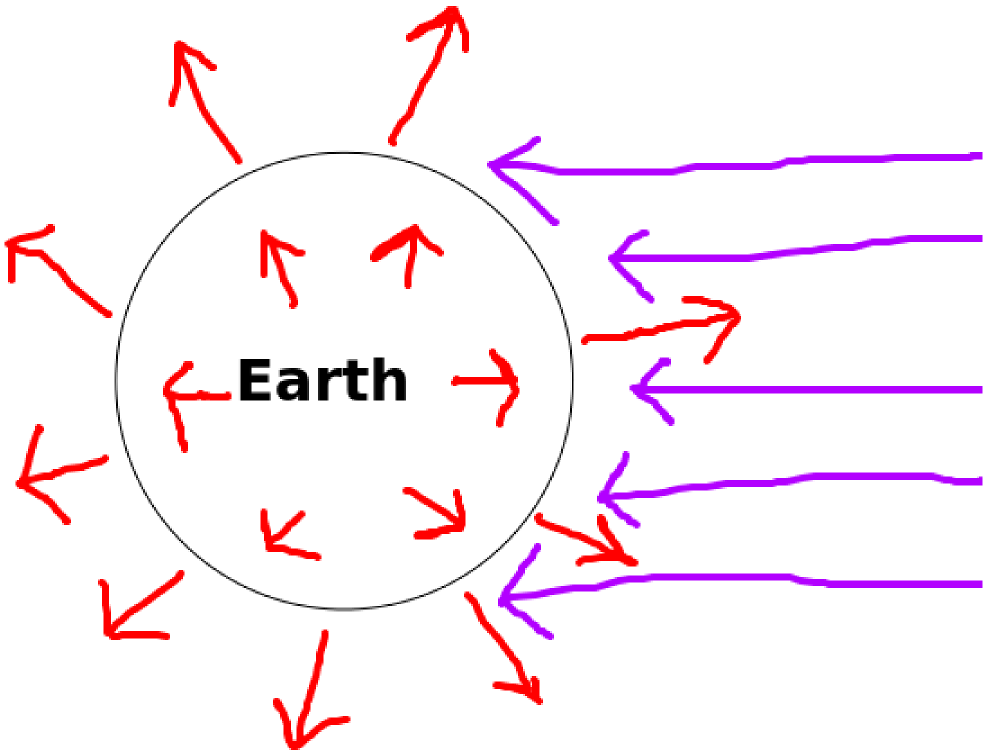

Image 4.2 - Shortwave radiation strikes the cross-section of the Earth (circular), but longwave radiation is emitted across the entire surface (spherical).

The adjustment for this becomes, firstly, multiply the incoming shortwave flux by the Earth's cross-section πr^2 and then divide that amount by the Earth's surface area $4\pi r^2$. Which is to say, divide the incoming flux by 4 to find the outgoing flux under this model - $\sim$321 W/m^2.

Our final step is to repeat the Stefan-Boltzmann law in reverse - compute the Earth's surface temperature under this model using our calculated flux. This computation yields a temperature of $\sim$274.3 K under this model.

You might have noticed that the temperature we computed is substantially lower than average real-world temperatures on Earth (274.3 Kelvin is approximately the freezing point of water). This is because, although the Sun is fairly close to being an ideal blackbody, the Earth is much more complex. We will examine the effects of some of these complexities in the next section, although a full mathematical treatment of these additional effects will be left as an exercise for the reader.

5.2 Albedo, Absorption, Reemission, and Radiative Forcing

An ideal blackbody absorbs all light that hits it, but in reality different parts of the Earth vary greatly in how much shortwave radiation is absorbed - some of it will instead be reflected back to space. The proportion of shortwave radiation that an object reflects is called albedo. Some objects, like open ocean, or dense vegetation, have low albedo (absorb most shortwave radiation, reflecting only a small amount), while things like ice, snow, and some types of clouds tend to have higher albedo, sometimes as much as 0.8 (meaning they reflect 80% of shortwave radiation). Notice how high albedo objects tend to be lighter while low albedo objects tend to be darker - this is a result of approximately half of the Sun's radiation being in the rather narrow visible portion of the electromagnetic spectrum, so there is a good absolute correspondence between how much light objects reflect across the entire spectrum, and how much they reflect in the visible range. Reflection of shortwave radiation into space is a cooling effect, reducing the Earth's equilibrium temperature, because at equilibrium it is not radiating as much energy if some of it has already been reflected instead of absorbed.

The other major deviation from the blackbody model is that the Earth is not a single-layered object, but in fact has a complex atmosphere which itself interacts with shortwave and longwave radiation. The atmosphere has its own albedo, dependent on thickness and composition, but the strong warming effect that pushes Earth's temperature upward is the result of absorption of longwave and reemission thereof by molecules in the atmosphere, i.e. greenhouse gases. When greenhouse gases absorb longwave radiation emitted by the Earth en route to space, they reemit in all directions, so the net outbound radiation decreases, and the Earth must be hotter to compensate (i.e. it would be out of equilibrium at the lower temperature, and heats up from the downward-reemitted radiation until it gets hot enough to balance out into equilibrium).

Frequently, changes in the Earth's equilibrium temperature form a feedback loop, where the temperature change affects aspects of the Earth's surface or atmosphere, which then affect the strength of future changes. For example, sea ice has a high albedo, but becomes less common when the Earth's temperature increases (as the marginal temperature shift above the freezing point of water melts some of it), which produces a positive feedback loop - rising temperature melts ice, which makes the Earth absorb more energy and warm a bit more, and so on. The opposite phenomenon occurs with cloud cover - warming causes more cloud formation (via evaporation from the oceans), and increasing cloud cover increases the Earth's average albedo, which dampens the rising temperatures somewhat.

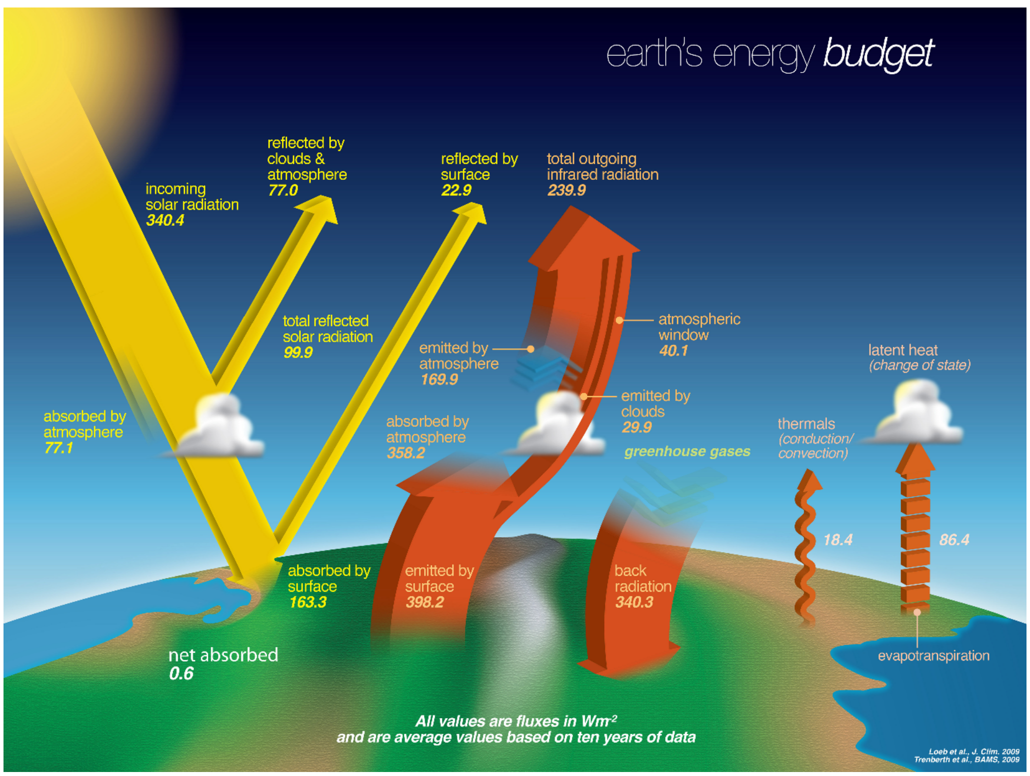

Image 4.3 - Summary of the Earth's energy balance.

There are many facets of these interactions that affect the Earth's energy balance. Image 4.3 summarizes some of the general inbound and outbound flows and interactions. Notice how the values shown here deviate from the computed numbers of the previous subsection due to the simplicity of that initial model. Understanding the factors influencing the Earth's energy balance is a core component for modeling and predicting climate change.