Module 1.1

Slide 1: In this short module, we will cover the different types of EM radiation used for remote sensing and review part of the US radio frequency spectrum.

Slide 2: EM radiation occurs in a wide range of wavelengths and corresponding frequencies. In the gray box, different types of radiation are labeled surrounding their corresponding wavelength, on the top axis, and frequency, on the bottom axis. In this course, we will discuss applications in the microwave, infrared, and visible part of the EM spectrum. First, we will discuss passive instruments in the visible and infrared, then passive microwave, then finally active sensing instruments such as radar. The US Radio Spectrum, outlined on the next slide, is bounded by the red lines.

Slide 3: This busy figure shows different allocations within the radio and microwave part of the spectrum. Each color represents a different type of allocation, which you can see by zooming in on the figure in your set of slides for this video. The full text table can be found at the link below the figure. Broad swaths are reserved for radio and TV broadcasting, particularly at less than 800 MHz. At higher frequencies, allocations tend to be narrower. A few meteorological applications that we will cover in this class are circled in white: S-band, C-band, and X-band radar.

Slide 4: The next figure is zoomed in on the second to bottom row on the previous slide. Again, C-band and X-band are circled, and black arrows point to other meteorological and earth exploration satellite allocations. Over the course of this class, we will discuss why certain bands are preferred for various applications.

Module 1.2

Slide 1: This module covers a few fundamentals of absorption and scattering in the atmosphere. We will discuss transmissivity of the atmosphere and atmospheric windows.

Slide 2: Satellite instruments measure radiance, which is related to the intensity of radiation at a certain wavelength, at the top of the atmosphere. Sources of the radiation include emission by objects on Earth or scattering of radiation into a path directed toward the satellite. Processes that reduce the radiance observed by a satellite are absorption and scattering away from the path toward the satellite.

Slide 3: The graphic describes an example of radiative transfer through a cloudy atmosphere. Red lines indicate sinks, and black lines denote sources. Radiation emitted by the ground, ocean, or the atmosphere below the cloud can either be absorbed by the cloud, or scattered by the cloud in any direction away from the satellite. However, the cloud emits a different amount of radiation at the same wavelength, and can scatter radiation initially not moving toward the satellite into the direction of the satellite. Suppose the radiance at the surface is L, and the radiance at the top of atmosphere is L_{top}. The difference, dL, is a sum of four terms: Increases due to emission and scattering into the path to the satellite, and decreases due to absorption and scattering away from the path to the satellite.

Slide 4: Atmospheric transmissivity is shown here as a function of wavelength. In this figure, the gray shaded area denotes how much radiation can pass through the entirety of an atmosphere with average temperature and moisture content for different wavelengths. Where the gray area reaches the top of the figure, the atmosphere is particularly transparent. This happens in particular in the visible part of the spectrum and in parts of the microwave. In the infrared, there are several bands—or ranges of wavelengths—in which the atmosphere is transparent. For example, the atmosphere is fairly transparent to 11-micron radiation. Such bands are known as **atmospheric windows**. Other bands are highly opaque. For example, at 7 microns, water vapor is a strong absorber. One would not want to use such a band for observing IR emissions from the surface, especially in locations where water vapor is often present.

Slide 5: One can also see the role of various common molecules in the atmosphere on transmissivity of radiation. Here, we zoom in on infrared wavelengths between 1 and 15 microns. Each row represents the transmissivity of an atmosphere containing an average amount of some molecule. For example, ozone, shown in the fourth row, is not transparent around 9.5 microns. This is because ozone absorbs radiation at this wavelength. The bottom row contains the summed effects of all atmospheric constituents. You can see that water vapor dominates the total transmittance at most wavelengths because the bottom two plots are quite similar. However, you can see some bands at which molecules other than water vapor are particularly important. For example, water vapor allows 9.5-micron radiation to pass, but ozone does not. This means that the atmospheric window spanning approximately 8–12 microns excludes the narrow ozone band. Carbon dioxide causes absorption above 13 microns and at some other wavelength ranges, such as around 4.2 microns. The effects of well-mixed molecules, like carbon dioxide, will be similar around the world. However, the effects of other molecules, like water vapor, are heavily dependent upon how much of that molecule is actually present. If no water vapor is present, then it will not affect the transmittance of the atmosphere. We will use this fact as a powerful tool to use passive radiation detection in multiple bands to tell us about water vapor concentration in the atmosphere.

Slide 6: So far, we have talked about the effects of Earth's atmosphere on **absorption** of radiation. However, as we saw before, scattering can act as both a source *and* a sink of radiation detected by a satellite. On the abscissa of this plot is wavelength of radiation, and on the ordinate is the radius of different scatterers in the atmosphere. We define a **size parameter**, which is 2-pi-r divided by lambda, or a ratio of the circumference of a hypothetical spherical scatterer to the wavelength of radiation. Rayleigh scattering occurs when the size parameter is small. Such scattering interactions tend to disperse radiation isotropically, or equally in all directions. As the size parameter approaches and exceeds 1, we enter the Mie scattering regime, which results in more complicated scattering interactions that are dominantly forward scattering. You can see the chi = 1 line in solid black running diagonally here. The slope of these lines means that shorter wavelengths are more efficiently scattered in the atmosphere. For example, visible light, which is centered around 0.6 microns is scattered by smoke, dust, and haze, but longer wavelengths, such as thermal IR radiation around 10 microns, or microwave are not scattered as much. Blue light is also more efficiently scattered than longer wavelength red light, which largely explains why the sky appears blue. For remote sensing purposes, this means that we are not particularly concerned with scattering in the infrared by the clear-air atmosphere. The primary source of radiation at the top of atmosphere for IR is emission by the surface or atmosphere, and the primary sink is absorption along the path to the satellite. Visible light, however, depends primarily on scattering to be detected. For example, for a satellite to detect a cloud, visible light must be reflected, or scattered, off the water or ice hydrometeors that make up the cloud. As you can see in this plot, cloud drops and rain drops are very efficient scatterers of visible light. Another example is that small aerosols are transparent to IR radiation, for example those resulting from biomass burning which are common over the West Pacific and areas like the South China Sea, reflect blue light more than red light. We will discuss scattering in more detail when we cover radar later in the quarter.

Module 1.3

Slide 1: In this video, we will review some basics of Planck's Law, which we will refer back to several times throughout the quarter when discussing passive remote sensing.

Slide 2: The **Planck function** describes the EM radiation emitted by a **blackbody** in thermal equilibrium at some temperature *T*. Remember from your radiative transfer class that a blackbody is a perfect absorber and a perfect emitter. Therefore, it obeys Kirchhoff's Law, which states that emissivity of an object Is exactly equal to its absorptivity, neither of which can exceed 1. We will approximate that many objects in the atmosphere are blackbodies. The Planck function has a complicated looking mathematical structure that contains an inverse relationship with wavelength, but also an exponential in the denominator that includes both wavelength and temperature. The structure of the Planck function, plotted on a log-log plot is shown at right, with Planck radiance shown as a function of wavelength. The different curves show the Planck radiance for blackbodies with temperatures ranging from 3K to 10000K. The sun emits around 5800K, and Earth emits at around 300K. You can see that the Planck radiance maximizes at smaller wavelengths for higher temperatures. For example, a blackbody at 5800K has a maximum emission in the visible light part of the spectrum. The 300K blackbody emits the most at wavelengths a little above 10 microns but does not emit detectable visible light. The wavelength of maximum emission can be described by Wien's Law, which simply states, that the wavelength of maximum emission is equal to a constant divided by the blackbody's temperature. Plug in a few different values into Wien's Law and see what you get.

Slide 3: In this figure, the solar spectral irradiance at the top of Earth's atmosphere is plotted by the top solid black line. Note that this figure is plotted on a linear scale instead of on a log-log plot. You can see how this compares to what would be expected from a 6000K blackbody, shown by the dashed line, which indicates a wavelength of maximum emission around 500 nm. This means that the sun emits the most in the blue visible light part of the EM spectrum. The Planck curve for the 6000K body and the TOA solar radiation goes to zero quickly as wavelengths go to 4 microns. For remote sensing applications, this means that we only try to detect reflected solar radiation at wavelengths less than about 4 microns.

Slide 4: The actual solar irradiance at the Earth's surface for an atmosphere with average water vapor content is depicted by the bottom of the shaded region. Molecular oxygen and water vapor absorb near-infrared solar radiation in various bands. Some of the bands are denoted by the red circles. If you look at the different curves around 1.4 microns, you'll see that the top of atmosphere irradiance is about 400 W/m^2/micron, but the surface irradiance is 0. You wouldn't want to use a 1.4-micron wavelength on a satellite to gather information about reflection from the ground!

Slide 5: Finally, scattering reduces how much radiation reaches the surface compared to the TOA. The difference between the top of the shaded region and the top black line indicates the sink—as a function of wavelength—caused by scattering. The blue and red lines respectively represent scattering of blue and red visible light. As discussed in the previous video, blue light is scattered more than red light by the atmosphere. Near IR radiation is not scattered comparably little as seen by the small difference between lines at longer wavelengths.

Slide 6: We can also look at normalized Planck curves describing the downwelling solar irradiance at the surface and upwelling terrestrial irradiance at the TOA on the same plot. On the top row, the red line represents a representation of downwelling surface radiation from the sun, approximated as a 5525K blackbody. The purple, blue, and black lines represent irradiance corresponding with Planck emission curves for Earth at 210K, 260K, and 310K. They represent the TOA irradiance for a perfectly transparent atmosphere. Note that the displayed curves for irradiance are *normalized*. Following Planck's Law, blackbody *radiance* from the warmer body is greater at *all* wavelengths. The shaded red and blue regions indicate the actual irradiance as a function of wavelength. As shown on the previous slides, the solar irradiance is reduced by scattering and absorption in certain bands. The bottom panels plot total extinction of radiation, or the combined loss due to scattering and absorption, of the atmosphere, broken down into contributions by various molecules. This is just 1 minus direct transmittance of the atmosphere. Water vapor absorption bands are clearly seen in the solar part of the spectrum. The blue shaded region represents the IR atmospheric window around 8–12 microns, where water vapor is not a strong absorber or emitter. The contributions of other molecules can also be seen and are important to consider for various applications that we will soon discuss.

Module 1.4

Slide 1: In this video, we will review some of the basic mathematics behind extinction, which means absorption or scattering, of radiation in the atmosphere. We will see various forms of Schwarzchild's equation, which describes how radiance changes through a layer of the atmosphere. We will also cover the concepts of direct transmissivity and optical depth.

Slide 2: Let's return to our idealized atmosphere from a couple of modules prior. The radiance observed at the satellite is whatever comes from the surface plus the sources along the path to the satellite, caused by emission and scattering, minus the sinks, caused by absorption and scattering.

Slide 3: Suppose you are interested in the change in radiance, or *dL*, and any point that we will call **X**, which in this case we have put in the cloud. The change in radiance is a function of location and the direction of the path to the satellite, **r**. A, B, C, and D simply represent sources and sinks of radiation. In its most simplistic form, this is Schwarzchild's equation; however, A, B, C, and D aren't particularly informative. We will expand on this a little bit using the second equation shown here. Now *dL* is displayed as only two terms, the first of which is the total sink, while the second is the total source.

Slide 4: L-sub-lambda here represents the initial radiance at point **X** in direction **r**. Sigma-e represents the volume extinction coefficient at the point, which is the sum of the absorption and scattering coefficients. Each of these cofficients has units of inverse length. We can also define a **single-scattering albedo**, which is the ratio of the volume scattering coefficient to the volume extinction coefficient. It describes the probability of an interaction between a photon and a potential scatterer as being a scattering interaction instead of an absorption. The single-scattering albedo is low for wavelengths that are not scattered in the medium through which they propagate—for example microwaves through clear, dry air.

Slide 5: *J* represents the sources of radiation at that point toward the satellite. It consists of a source from thermal emissions and a source from scattering in a path toward the satellite.

Slide 6: The thermal emissions are simply described as sigma-a multiplied by the Planck radiance at point **X**, which is defined by the temperature at **X**. Note also that **Kirchhoff's Law** states that an object in thermodynamic equilibrium is an equally good emitter and absorber at the same wavelength. The coefficient related to emission will be denoted as epsilon. If it were 1, the body is a blackbody at the wavelength specified. Perfect blackbodies do not exist, but planets and stars can often be approximated as blackbodies, at least for the purposes of describing them in class. Note that objects can have epsilon that varies as a function of wavelength. If an object is a perfect emitter at some wavelength, it will also absorb all radiation incident upon it at the same wavelength. Here, I have written that the volume absorption coefficient is equal in magnitude to the emissivity. What is strictly true is that absorptance, which is unitless unlike the volume absorption coefficient, is equal to emissivity. For the purposes of this class, we will simply assume that sigma-a is defined over a unit path length and that its magnitude equals the absorptance coefficient, which like emissivity, cannot exceed 1. We will subsequently use sigma-a and epsilon interchangeably depending on whether we want to describe emission or absorption.

Slide 7: The source associated with scattering looks a bit more complicated. It is a function of the radiance at point **X** moving in any direction **r’**. The direction **r**, remember, is the path to the satellite. Gamma is just a scattering phase function that incorporates the probability of radiation traveling in direction **r’** getting scattered into direction **r**. Then we integrate over 4-pi steradians, the solid angle subtended by a sphere surrounding point **X**, through all directions **r’**, which is collectively denoted as upper-case omega’. To summarize the past few slides, each of the two terms seen in the equation on the left have two terms within them. We can further expand on this form of Schwarzchild’s equation and will do so in the next module for various idealized radiative transfer scenarios.

Slide 8: Before moving on, we should define a couple of very important concepts. The first is **optical depth**, which is also called optical thickness. I may use the two terms interchangeably throughout the course. I cannot stress one point enough: Optical depth is not actually a physical depth. It is a unitless variable that s how much absorption and scattering of radiation at some specified wavelength occurs along a path. The mathematical definition is the integral of the volume extinction coefficient integrated over some path. We will use small delta to denote optical thickness. What is displayed here is the **vertical path optical depth**, which describes the optical depth along a perfectly vertical path. If we integrate this from 0 to the top of the atmosphere, we compute the optical depth of the atmosphere. Another important concept is the **direct transmittance**. It exponentially decays as vertical path optical depth increases, and it is also dependent upon the angle off the vertical that describes the direction of the path. We will see some drawings of this next to help you visualize this. That angle is described as theta, and we will denote mu as the cosine of theta. If theta is 0, meaning a vertical path, then the direct transmittance is just e to the negative vertical path optical depth, and you are left with the direct transmittance (or transmissivity) of the atmosphere, which is what we discussed in previous modules. Finally, note the general expression of path optical depth, where the path is denoted by *s* instead of the vertical coordinate *z*. Delta as a function of s along a path from s_1 to s_2 is just delta as a function of z through the layer transited by the path from s_1 to s_2 divided by mu. With all those words done now, let’s look at some pictures. In the next module, we will cover more related to optical depth along different paths, so you may find coming back to this slide useful for understanding what we cover then.

Slide 9: Let’s return to our idealized atmosphere with a cloud. The ground has radiance L. For a path straight up, the angle theta is zero. Whatever the value of sigma-e at various altitudes, its integral will always increase as radiation moves upward from the ground, since sigma-e must be positive.

Slide 10: The optical depth along this path is just the vertical path optical depth, or the integral of sigma-e from the surface to the top of the atmosphere. Here, we are denoted the ground as z’=0 and the top of atmosphere as z’=z. Then we integrate over z’, which is our vertical coordinate.

Slide 11: We can also compute the optical depth of a layer that isn’t the entire depth of the atmosphere. For example, suppose we want to know the optical depth of the cloud layer. Then we simply integrate from the bottom of the cloud to the top of the cloud.

Slide 12: The path that radiation travels along, however, doesn’t have to be vertical. Suppose we want to know the optical depth along a path from the same point on the ground to the satellite in this drawing. In this case, we have to consider the angle theta. The path along the solid line to the satellite is longer than the dashed line that points straight upward, so the optical depth (assuming that sigma-e is horizontally homogenous) will be larger along the solid line. In this case, we can say that the ground is point s_1 and the satellite is point s_2. The direct transmittance along this longer path will also be lower than the transmissivity along a vertical path. Of course, you can imagine, that the optical depth might be very different dependent upon whether a cloud is present. It is also very sensitive to the concentration of atmospheric constituents, such as water vapor. The various absorptivity of water vapor and other molecules as various wavelengths explains why some wavelength ranges are atmospheric windows on Earth and others are not. We will continue to build upon these concepts over the next few modules and begin to apply them as we start discussing particular remote sensing instruments.

Module 1.5

Slide 1: In this lecture, we will describe a few idealized cases of Schwarzchild's equation, including the case described by Beer's Law. Recall the general from of Schwarzchild's equation, shown here at the bottom. We will use this as a starting point for discussion.

Slide 2: Consider again an atmosphere with optical depth that varies with height. From the point of view of a satellite, the optical depth at the top of the atmosphere is 0, and the optical depth of atmosphere in this example is delta-sub-t. We will use coordinates of optical depth as a proxy for height coordinates. Again, we know that optical depth is larger through the same atmosphere along a slant path than along a vertical path, and the two are related to each other though the cosine of the angle that the path to a satellite makes with the vertical path from the surface.

Slide 3: Recall the definition of optical thickness and differentiate both sides of that equation, and you get what you see here. Doing some simple algebra will yield the following equation that is just the Schwarzchild's equation with the possibility of a slant path included. Wherever you see phi, it just represents the azimuthal angle for the path to the satellite. For considering idealized cases though, we will just consider a two-dimensional space like that shown here.

Slide 4: Next, suppose you want to describe *dL* over several discrete small layers from the surface to the top of atmosphere. One such box is shown by the black box at height z. Radiation is emitted, absorbed, and scattered by molecules in the box. The radiance at the top of the atmosphere (at delta = 0) is equal to the radiance at the surface, reduced by the direct transmissivity of the entire atmosphere along the path length, plus the integral of the sources in these several discrete boxes along the path. However, sources of radiation are then subject to extinction further up the path, so only the direct transmittance of the layer above the box is considered in the integral. In this example, we're simply treating the integral by conceptually breaking up the atmosphere into many discrete layers.

Slide 5: Let's begin to consider some idealized cases. First, let's assume that no sources of radiation exist along the path. That means that *J* goes to zero. We can do some simple algebra to show that radiance at the end of the path, at s_1, is equal to radiance at the beginning of the path, at s, reduced by the direct transmittance. This looks like the first term in the equation from the previous slide. This relationship is known as the Beer-Lambert-Bouguer Law, also known as Beer's Law.

Slide 6: If we consider an idealized situation where there is no scattering *into* the beam and the Planck radiance at the wavelength of interest is approximately zero, the radiance at the top of the atmosphere is simply radiance at the bottom of the atmosphere, reduced by the direct transmittance, here also accounting for the angle even though the figure illustrates a vertical path length. If we consider the vertical path, an optical depth of 1 corresponds to a direct transmissivity of about 37%. High optical depths correspond to very low transmissivity. An optical depth of about 4.5 corresponds to direct transmissivity of 1%. Such optical depths are quite common at wavelengths that are efficiently absorbed or scattered. For example, clouds are highly opaque to visible light, and they commonly have optical thickness of greater than 1.

Slide 7: Let's look at a few examples. In this example, we are considering an atmosphere with three different profiles of aerosols or some other scattering/absorbing objects. Suppose radiance out of an ocean is 1000 W m^{-2} sr^{-1}. There are no sources of radiation along the slantwise path indicated in each panel. The first atmosphere, on the left, is a well-mixed homogenous atmosphere. The second atmosphere, at middle, has more aerosols in the lower troposphere than aloft. The third atmosphere is one that contains a layer of aerosols. It is possible for all three atmospheres to have the same direct transmittance. While the optical depth of the layer from z to z+dz varies across the three examples, and is largest in the third atmosphere, the optical depth of a low-tropospheric layer is largest in the second example, and largest in the upper troposphere in the first example. Thus, the optical depths of the entire atmospheres could feasibly all be the same. Let's suppose they are, and that the radiance at the top of each atmosphere is 400 W m^{-2} sr^{-1}. What is the optical depth of the atmosphere? Pause here, taking a little time before advancing to see if you can figure out how to calculate the optical depth.

Slide 8: You should get something close to 0.79. Divide 400 by 1000 to get 0.4. Take the negative natural log of 0.4 and multiply by the cosine of 30 degrees.

Slide 9: Consider the next simple case. Here we will allow emission, but not scattering, to be a source of radiation along the path. This is primarily the case for infrared emissions detected by satellites. Now *J* is not zero. Instead it is proportional to the Planck radiance multiplied by the emissivity of the surface. The Planck radiance is determined by the temperature of the surface. We can take the general form of the Schwarzchild's equation and substitute it for values related to the Planck radiance. The surface radiance can be replaced as shown in red, and the radiance along the path, the source, is substituted in for J over sigma-e in the integral.

Slide 10: Again, the solution we end up with says that radiance at the top of the atmosphere is equal to radiance at the surface reduced by transmittance, and added to that, all the sources along the path to the satellite, themselves reduced by the transmissivity of whatever layer above them through which they must pass.

Slide 11: What does this mean for remote sensing? Recall our discussion of atmospheric windows. Because of water vapor, carbon dioxide, ozone, and some other molecules, the atmosphere is opaque to radiation in certain wavelengths and transparent to it in others. Ask yourself, in atmospheric windows, such as those near 10 microns, is the blue Term 1 or the red Term 2 larger? What about in bands such as those centered on the water vapor absorption band near 7 microns? Term 1 is much larger in atmospheric windows because the direct transmissivity of the atmosphere is high, and most of the radiation emitted by the surface reaches the top of the atmosphere.

Slide 12: Finally, we will consider the case in which scattering Is the only source. A realistic example of this is solar radiation, which is not emitted. For example, visible satellite imagery sees radiation scattered off of objects on the surface or in the atmosphere. For simplicity, we will assume that single-scattering dominates. In other words, we'll assume that once radiation has been scattered one time, it will not be scattered again. This is not really a realistic assumption, but it is effective for explaining the basic concept of how scattering is incorporated into Schwarzchild's equation. In this case, J_{th} is zero, and we'll also assume that no radiation is emitted by the surface. Suppose radiance from the sun is described by L, with angles theta_0 and phi_0 related to the angle of incoming radiation. After radiation is scattered, it exits the atmosphere at some new angle theta and phi. Thus, at a given point in the atmosphere, the radiation at an arbitrary point **X** is the incoming solar radiation reduced by the transmissivity of the layer, which is controlled in this case by how much scattering occurs above point **X**. The radiance at a point along a path is then an integral of a complicated looking phase function, gamma, multiplied by the incoming radiance reduced by transmissivity of the layer above. The phase function can be expressed as a function of scattering angle, psi-sub-s, and again we integrate over the 4-pi steradians subtended by a sphere. In this class, we will not try to compute or use any of these phase functions, but I do want to at least skim the surface on how they can be expressed as a source or sink of radiation to a satellite.

Slide 13: We can start with our general form of the Schwarzchild's equation and plug our new expression for the source term in. Note here that we have switched the sign convention for the vertical coordinate, so that L-sub-t denotes radiation at the top of the atmosphere outbound, and L-0 denotes inbound radiation that is scattered off the surface. If we assume that the atmosphere is homogenous, we end up with the expressed at the bottom. It says that top of atmosphere radiance is equal to radiance scattered off the surface toward the satellite reduced by the direct transmittance of the entire atmosphere along a slant path plus any additional radiation scattered by the atmosphere into the direction of the satellite. Note that we have replaced sigma-s over sigma-e with the single scattering albedo, and that two exponential decay terms appear in purple: One for radiation on the way down, and another for radiation on the way up. Although we won't do so here, we can ultimately combine the various idealized cases to include absorption and scattering. Solar radiation is primarily scattered to satellites, while infrared radiation detected by satellites is primarily emitted. However, where there is overlap with solar and terrestrial emissions—around 4 microns—both must be considered. Additionally, in the microwave part of the EM spectrum, we will have to consider both absorption and scattering.

Module 1.6:

Slide 1: In this module, we will discuss the orbits of satellites in medium to high-Earth orbit, including geostationary satellites, which are very important for global meteorological observations and a focus of the second Lecture Series that will follow.

Slide 2: The orbit of a satellite is important primarily because it controls the area that we can see from the satellite and also puts a physical constraint on the spatial resolution that we can achieve. It also impacts the projection of a satellite image. In the python code for the first lab, you can actually see where the map projection for the data must be specified. Particular attributes of an orbit determine how much area a single satellite can cover, how high a latitude it can take observations, or what the return time is for a satellite at a given point on Earth.

Slide 3: An orbit is defined by a few characteristics. The first is height. We will discuss orbits in three categories of heights: Geosynchronous orbit, near 35,786 km above surface; which is where satellite platforms like GOES, Himawari, Meteosat, INSAT, or Fengyun operate. Mid-Earth orbit is essentially anything below geosynchronous orbit and above about 2,000 km. Semi-synchronous orbits are examples of mid-Earth orbits. Low-earth orbits are less than 2,000 km above the surface, and most that we will talk about in this class are under 1,000 km in altitude. The majority of satellite systems we will discuss in this class are in low-Earth orbit.

Slide 4: The speed of an object in orbit is controlled by Earth's gravity and the height of the orbit. Some typical values of geosynchronous and low-earth orbits are shown here alongside the corresponding velocity and how frequently the satellite completes one orbit. Based on Newton's laws of motion, the gravitational force between a satellite and the planet is inversely proportional to the distance squared between their centers of mass. A geosynchronous orbit is completed once daily. Satellites in low-earth orbit are at lower altitude, move more than twice as fast, and complete a full orbit much more frequently.

Slide 5: Eccentricity is the next important parameter used to describe an orbit. It describes the shape of the orbit. It is denoted here by epsilon. Eccentricity of zero describes a perfectly circular orbit, while higher eccentricities describe elliptical orbits with foci that are increasingly far apart, and one of which is always Earth. The orbits for most of the platforms we will discuss in this class have low eccentricity, meaning they are near-circular.

Slide 6: A couple of additional terms come up when discussing elliptical orbits: apogee and perigee. Apogee refers to the point in an orbit that is farthest from the major focus, which is Earth for remote sensing satellites. Perigee is the point of closest approach to Earth. Other orbits have these parameters too. For example, Earth's orbit around the sun is not perfectly circular. It reaches apogee (apihelion) in July and perigee (perihelion) in January.

Slide 7: An orbit is also defined by its inclination. The inclination describes the angle off the equator made by the sub-satellite path of an orbit. Imagine if you drew a line in the ground pointing straight down from a satellite and drug it along the ground as the satellite moved. That would be the sub-satellite point. We call the viewing geometry of the satellite pointing straight down from the satellite **nadir**. An outline of one hypothetical satellite track is shown by the red line in this figure. The angle denoted by the black arc is the orbital inclination. In this example, it looks to be about 45 degrees. Geostationary orbits have inclination, as we will soon see, of 0 degrees. An orbit directly over the poles would have an inclination of 90 degrees.

Slide 8: A special type of low-earth orbit, sun-synchronous orbit, has an inclination of around 98 degrees, which is denoted by the blue line. That means that the highest latitude over which it will pass is 82 degrees, although onboard instruments may be able to sample off-nadir at higher latitudes. Orbits with inclination larger than 90 degrees are in retrograde orbit, which means they appear to move from east to west on successive orbits. Prograde low-earth orbits are those with inclination less than 90°.

Slide 9: As stated before, the orbits for most satellites we will discuss in this class are close to circular. However, there are several factors that make the orbit slightly elliptical or require occasional correction to maintain the desired orbit. The most obvious is that the distribution of mass on Earth is not spherical. The Earth's geoid has numerous fluctuations in space—which also becomes important to consider when collecting measurements with an altimeter. Other bodies—particularly the sun and moon—also exert gravitational force on satellites. Especially for satellites in low-Earth orbit, drag and lift are dependent upon properties of the satellite platform itself and must be accounted for because they cause the orbital velocity to be a little different than that predicted by Newtonian mechanics. Other processes associated with solar radiation and Earth's EM field can cause higher order minor alterations to orbit that can vary with time. Occasionally, corrections must be made to an orbit by using fuel to thrust a satellite platform in a direction that increases or decreases its forward velocity in order to maintain a stable, desired orbit for observations that are consistent and high-quality for many years.

Slide 10: Next, we'll look at geosynchronous orbits—and in particular, the special case of geostationary orbits—in more detail.

Slide 11: A geosynchronous orbit is any orbit at the height of about 35,786 km above mean sea level. It can have any inclination, and a variety of combinations of inclinations and eccentricities can be used to create various orbital patterns. The main characteristic of geosynchronous orbits is that they return to the same zenith, or overhead, point every 24 hours.

Slide 12: A geostationary orbit is a special case of geosynchronous orbit as denoted by the simple Euler diagram shown here. Its inclination is zero, meaning it is neither in prograde or retrograde. This means that a geostationary orbiter remains over the same sub-satellite point all the time. A geostationary satellite is therefore often defined by the longitude of its nadir viewing angle. In other words, it views the same scene continuously. This is extremely useful for making observations of Earth with high temporal resolution. The entire second lecture series is devoted to interpreting measurements from advanced imagers aboard the United States GOES and Japanese Himawari weather satellites.

Slide 13: The velocity and height of a geostationary orbit can be determined analytically easily by using Newton’s laws of motion. Remember Newton’s second law, which states that the force needed to accelerate an object is proportional to its mass. Perhaps you recall from a university course in kinematics, that centripetal acceleration is velocity-squared of an object divided by its orbital radius, the latter which is the distance from the center of rotation—in this case Earth’s center of gravity. Note the expression for gravitational force between two bodies, which is proportional to the product of the masses of the two objects, and as stated earlier, inversely proportional to the orbital radius squared. If we equate the two to each other, then the mass of the satellite becomes inconsequential to its orbital velocity, although a large aerodynamically “sticky” satellite may have more drag and make its orbital velocity a little slower than that predicted by the Newtonian mechanics laid out here that neglect friction. We know that the orbital period is just the distance of the orbit, which is approximated here as the circumference of a circle, divided by the orbital velocity. Then do some algebra to solve for the orbital period, or given the period, solve for the radius. Try this and see if you can mathematically show what the height of geostationary orbit is.

Slide 14: An example of the view from a geostationary satellite is shown in this image, taken from GOES-16. As you can see, the entire disk of the Earth is visible. If you were able to zoom in and look closely, you could see that the spatial resolution is higher at nadir than it is at the edges of the disk. An image like this can be collected currently every 15 minutes, with imagery over smaller sectors collected more frequently.

Slide 15: As mentioned, one of the benefits of geostationary orbit is its large spatial coverage that is continuous. However, the spatial resolution is not very high. The highest resolution of any GOES data at nadir is 500 meters by 500 meters. For comparison, we will see later in the course some examples of LEO satellites that can achieve resolution on the order of a few centimeters. The distance also makes active sensing impractical. Furthermore, as seen in the previous image, polar regions are not well observed, and the high latitudes that are observed are done so at coarse spatial resolution. Finally, geostationary satellites experience temporary outages during eclipses.

Slide 16: We will finish this module by briefly discussing a couple of medium Earth orbits, the semi-synchronous orbit and the Molniya orbit.

Slide 17: A semi-synchronous orbit is situated at about 20,200 km above mean sea level. The image shown here isn’t to scale; the Earth is way too large compared to the orbits, but it just illustrates that the semi-synchronous orbit is always inside a geosynchronous orbit with the same inclination.

Slide 18: Satellites in semi-synchronous orbit move more quickly than those in geostationary orbit and complete one revolution every 12 hours. The primary use for semi-synchronous orbit discussed in this class is data derived from GPS satellite constellations. GPS operates at an inclination of about 55°, and GPS radio occultation is used to derive profiles of temperature and humidity in the atmosphere.

Slide 19: Finally, the Molniya orbit is seldom used for Earth science purposes but is highly useful for viewing or transmitting to/from high latitudes. This orbit is highly eccentric and has an inclination of about 63.4 degrees. It has a period of about 12 hours and an apogee of nearly 40,000 km. The orbital velocity varies with distance from Earth, increasing as the satellite passes closest to Earth. This means that the sub-satellite point of a platform in Molniya orbit is usually over high-latitudes. This is useful for purposes such as communications at high latitudes, where geostationary platforms are not very effective. An example of the sub-satellite track of a Molniya orbit is shown at bottom left. Because the orbital period is 12 hours, successive orbits view opposite sides of the globe. For Earth science purposes, a constellation of satellites in such an orbit could be useful for viewing weather or sea ice, for example, at high latitudes. In the next module, we will discuss low-Earth orbits in more detail.

Slide 20: Lagrange points are special locations associated with two bodies each having gravitational pull on an object at those points. They are locations where the sum of the gravitational force toward both bodies (in this case the Sun and Earth) are balanced with the centripetal force. Shown in this diagram are the Lagrange points for the Earth-Sun system. They are denoted as L1 through L5. The white lines indicate the gravitational potential, and the blue line denotes the approximate Earth orbit around the sun. Red arrows denote gravitational potential toward a Lagrange point, and blue arrows denote potential falling away from a point. The Lagrange points L1, L2, and L3 are always on a line extending between and beyond the Sun and Earth. L4 and L5 are located at corners of an equilateral triangle with L4 and L5. You can almost think about the white lines as representing a topographic map. The “peaks” are relatively flat ridges at L4 and L5. In contrast, the Sun and Earth represents gravitational sinks. You can think of these as deep valleys or bowls. L3 is located at a low point between the ridges at L4 and L5. L1 is located at a col (or like a “gap” or “pass” if thinking in terms of navigation in the mountains) between the Sun and Earth—a high point in the gravitational potential between the two. L2 is located at another high point between Earth and space on the side opposite the Sun.

Slide 21: The points L1, L2, and L3 are unstable locations. For example, imagine you had a little ball position at point L1, and you gave it just a little nudge in the direction of the blue arrows. The ball would pick up speed as it rolled away from the point L1. The same is true of L2 and L3. However, at points L4 and L5, the Coriolis force causes a body trying to move away from L4 or L5 to move back inward toward those points. Therefore, natural “trojan” satellites are often found at the L4 and L5 points of various planets. The L1 and L2 Lagrange points are of particular interest for remote sensing. For example, the James Webb telescope, which is used for astronomy and looks away from the Sun, is located at the Earth-Sun L2 point—on the side of Earth away from the Sun. The Deep Space Climate Observatory (DSCOVR) is situated at the Earth-Sun L1 point, and houses instrumentation used for space “weather” characterization, a topic that will be discussed in a module in the next lecture series. Typically, sensors are not situated at the L3 point because the Sun would be located between the sensor and Earth, making sending signals between the satellite and Earth impractical. And sensors are often not located at L4 or L5 because so many natural objects are present near these locations and would pose increased hazard to the instrument platform.

Module 1.7:

Slide 1: We continue our discussion of orbits in this module by examining low-Earth orbits.

Slide 2: The map shown is a somewhat out-of-date representation of global surface weather stations on Earth. As expected, land masses are generally much more thoroughly observed than oceans, where calibrated observations are generally only collected on islands. Weather balloons are only regularly launched from land as well. Thus, the majority of Earth is not well represented by direct observations, which motivates the use of remote sensing to fill in observational gaps.

Slide 3: An example of a low-Earth orbit is shown here. It usually has a high inclination that allows it to pass over high latitudes and low latitudes. Contrast this with the geostationary orbit that has zero inclination. Geostationary orbiters provide fantastic temporal resolution but lack in spatial resolution and latitudinal coverage. Constellations of low-Earth orbit satellites, on the other hand, can provide data with high spatial resolution and at high or low latitudes.

Slide 4: A few terms are important to understand when discussing low-Earth orbits and data from satellites in such orbit. Since low-Earth orbit is roughly around the poles, a satellite moves from south-to-north in either the Western or Eastern Hemisphere, then from north-to-south in the other. The south-to-north track is called the “ascending orbit”, and the north-to-south track is called the “descending orbit”.

Slide 5: A sun-synchronous orbit is one with a particular inclination, about 98°, that causes the solar angle, or the angle in the sky where the sun is located relative to a viewer on Earth, to be about the same each time the satellite passes the equator. Such an orbit is in retrograde, meaning that the longitude of the sub-satellite point drifts westward with successive orbits. The solar angle is determined by the equatorial crossing time, which roughly describes the local time—again based on the solar angle—that the satellite passes over. This is typically designed to be between 10AM and 2PM in one hemisphere. If a satellite orbit is designed to be moving northward over the equator at 10:30AM, then we would say that the orbit is “daytime ascending” with a 10:30AM equatorial crossing time. A sun-synchronous orbit allows for direct comparison of observations collected in different parts of Earth because the solar angle is approximately the same at each longitude observed.

Slide 6: Not all low-Earth orbiting satellites are in sun-synchronous orbit, however. A high inclination orbit allows for nearly complete global coverage; however, some satellite platforms are designed with lower inclination if their missions do not involve observations of the poles. For example, the Tropical Rainfall Measurement Mission was at a 35° inclination and viewed only the subtropics and tropics. The replacement for TRMM, the Global Precipitation Measurement Core Observatory, has an inclination of 65° and can view higher latitudes than TRMM could. Since these orbits are not sun-synchronous, they can make observations of locations at various times of day.

Slide 7: An example of a sun-synchronous orbit is that of the Aqua satellite. An example of four and a half successive orbits is shown at bottom. It has a daytime ascending orbit with an equatorial crossing time of 1:30PM. The exact track is repeated every 16 days.

Slide 8: Satellite in low-Earth orbit have limited field of view compared to geostationary satellites. In this figure, the satellite might only see the atmosphere and surface within the two outer yellow dashed lines, with the center dashed line representing the sub-satellite point. This means that data is collected in swaths.

Slide 9: An example of the swaths is shown in a day's worth of data shown here. There are gaps in coverage at the poles because the satellite does not orbit directly over the pole. There are also gaps near the equator. The swath width for most instruments is usually too small for the field of view in successive orbits to overlap.

Slide 10: Similar gaps at the equator can be seen in these examples of MODIS imagery from the Aqua and Terra satellites. Aqua has a daytime ascending orbit, while Terra has a nighttime ascending orbit. The two collectively overlap in coverage but view scenes at different angles.

Slide 11: Satellite orbits can also be organized into constellations. The A-Train is an example of one such constellation of low-Earth orbiting satellites. As of 2020, the A-Train consists of four satellites that are able to provide observations of aerosol, clouds, radiative fluxes, and atmospheric thermodynamic profiles. They are all in sun-synchronous orbit at an altitude of 705 km and pass over the same point on Earth within minutes of each other to provide comprehensive measurements of several different atmospheric variables coincident in space and time. The C-train consists of two active sensors: CloudSat and CALIPSO, which used to be members of the A-Train. They have been moved to a lower orbit in preparation for eventual re-entry into Earth's atmosphere at the end of the lives of each. The C-train orbits at a lower altitude and therefore moves at slightly faster velocity. We will see some data from these two satellites near the end of this course.

To start out the course, we will discuss geostationary satellite data in the next series of modules.