Module 2.1

Slide 1: The next series of videos will delve into modern instruments aboard geostationary and polar-orbiting satellites as well as interpretation and application of the data we can collect from these instruments. In the first module, we will look into the instrumentation and wavelength bands, or channels, used on modern GOES and Himawari satellites.

Slide 2: The instrumentation on GOES serves multiple purposes. What we will discuss in this class are the nadir-pointing Advanced Baseline Imager, or ABI, and Global Lightning Mapper, or GLM. We'll discuss the ABI in much more detail and the GLM in a different module. The top panel shows the size of the solar charging panels relative to the main satellite bus. The ABI, however, is the passive remote sensing instrument that detects radiation in a variety of visible and IR wavelength bands. GLM is a passive instrument that detects emissions caused by the heating of the atmosphere during lightning. Two instruments are solar-pointing: SUVI is the solar ultraviolet imager; as its name suggests, it collects full-disk imagery of UV radiation emitted by the sun at high temporal resolution. EXIS is the Extreme Ultraviolet and X-Ray Irradiance Sensor. Observation of both UV and X-ray radiation can provide some lead time for geomagnetic storms that may impact electrical power grids, radio communication through changes to the magnetosphere or ionosphere. The Space Environment In-Situ Suite (SEISS) and magnetometer (MAG) respectively monitor ion fluxes and magnetic field variations at the location of the satellite, or *in situ*.

Slide 3: As just a reminder, the orbit is about 42,164 km from the center of the Earth, or an altitude of 35,786 km. The distance is shown to scale in this slide, although, obviously the satellite size is not shown to scale.

Slide 4: The ABI has 16 bands, or channels. This, and the presence of many instruments on other platforms in low-Earth orbit, eliminates the need for having a separate sounder that consists of multiple observation bands surrounding some absorption band for the purposes of deriving vertical profiles of water vapor or temperature. The ABI can take a full disk image every 5 minutes, but the default is to take one image every 15 minutes, and in addition, collect imagery over the continental US every 5 minutes and within one or two mesoscale regions every 30 to 60 seconds. The spatial, temporal, and spectral resolutions are all markedly improved compared to the previous generation of GOES satellites.

Slide 5: The 16 channels on the ABI are listed here. The Advanced Himawari Imager (AHI) has the same bands except it does not have a “cirrus” band at Channel 4 but has a “green” visible band at 510 nm. I won't discuss every single band right now, but you can use this as a handy reference. Generally, IR bands from Channels 7–16 have a 2 km instantaneous geometric field of view at nadir, while some of the shortwave bands, or Channels 1–6, have higher resolution. I will often use the terms shortwave and longwave to respectively describe radiation that is either emitted by the sun or emitted by Earth. Each channel observes the spectral radiance averaged over the width of each band, which is indicated in the second column. The higher resolution is possible because of the higher radiance at the satellite sensor in the reflected shortwave bands. In other words, the reflected solar radiation is greater than emitted terrestrial radiation within the same solid angle, which makes sense given the discrepancy in temperature between the sun and Earth. The width of the band also controls radiance at a satellite sensor; a wider band will detect more radiation; thereby requiring a smaller field of view to attain enough signal to process.

Slide 6: The following imagery shows a sample of full disk imagery for five different geostationary satellites. The first shows the disk viewable by GOES-16, or GOES-EAST, centered at 75 degrees west, which views the Atlantic and the Americas.

Slide 7: GOES-17, or GOES-WEST, at 137 degrees west, which views the western US plus the eastern Pacific.

Slide 8: Himawari-8 at 140 degrees east, which views the western Pacific and eastern Indian Ocean.

Slide 9: INSAT-3 at 82 degrees east, which views South Asia and the Indian Ocean.

Slide 9: Meteosat-8 at 42 degrees east, which views eastern Africa, the Middle East, and the western Indian Ocean.

Slide 10: And Meteosat-11 at the prime meridian, which views western Africa, Europe, and the eastern Atlantic. Not shown here are also INSAT-3DR, another Indian geostationary weather satellite, stationed near 74 degrees east; and Feng-yun 4A, the most recent Chinese geostationary weather satellite that is stationed near 105 degrees east and has instrumentation similar to that onboard GOES and Himawari. The current generation of Meteosat and INSAT have lower spectral and spatial resolution than the newer systems but are comparable to the previous generation of GOES. Together, the satellites provide global full disk coverage of clouds in the tropics and middle latitudes every 30 minutes. In the next modules, we'll focus on GOES imagery before moving on to low-Earth orbiting data from the Joint Polar Satellite System, which is a joint effort between Japan and the United States.

Module 2.2:

Slide 1: In this module, we will begin to dive more deeply into the bands on the GOES ABI and explain some of their various utilities.

Slide 2: We will start back at the table of 16 GOES channels. Each channel represents a passive observation collected by detecting radiance with the wavelength range denoted by the second column in the table.

Slide 3: The spectral response function, or SRF, describes the sensitivity of bands to radiation as a function of wavelength, and it has a maximum value of 1. These functions are shown here for the AHI shortwave bands, with visible light colored. On the top row, you see the spectral response functions plotted as black lines for the blue, green, and red bands. Take the blue band for example; the radiance observed for Channel 1 is influenced by radiation with wavelengths between about 450 nm and 490 nm. However, it is more influenced by radiation closer to 490 nm than on the lower end of the range. Channel 2 is evenly influenced by radiation with wavelengths between about 500 nm and 520 nm. The red band, Channel 3, is a wider band that is influenced most by radiation with wavelengths between about 600 nm and 680 nm. Radiation at wavelengths where the SRF is zero for a given band, do not affect radiance observed in that band.

Slide 4: The spectral response functions for ABI are plotted in blue (except for the two visible channels) in this figure. Shortwave channels 1 through 6 are shown on the top row, and the shortwave infrared channel 7 and exclusively longwave channels 7 through 16 are shown at bottom.

Slide 5: For shortwave radiation, we are observing photons that are scattered by the atmosphere, things in the atmosphere such as clouds, or by the surface or ocean. Radiation enters at some angle called the incidence angle and the angle with the ground that the path to a satellite makes is called the exitance angle, which is 90 degrees minus the angle used to calculate mu in the computation for slant path optical depth.

Slide 6: Our goal is to take two known quantities and derive something about a third unknown quantity. We know what solar irradiance is, and we know what the satellite senses because it regularly transmits data to Earth. This is an example of an inverse radiative transfer problem. We know the “answer”, or the radiance observed, and we are trying to learn some properties of the surface or atmosphere that cause scattering on a path to the satellite sensor.

Slide 7: Back to our plot of SRFs, the top panel denotes channels that will detect scattered solar radiation. You can see how the red band is a little wider than the blue band, for example. The faint gray line represents the direct transmittance of the atmosphere as a function of wavelength, with its axis on the right side of the plot. For the most part, the bands are located in parts of the EM spectrum for which Earth’s atmosphere is transparent. For example, about

80% of the radiation reflected off the surface in Band 1—centered on 470nm—will reach the ABI; the rest will be scattered. Nearly 100% of the approximately 1.6-micron radiation scattered off the surface will return back to space. Five of the six shortwave bands can readily pass through the atmosphere without being heavily absorbed. However, Channel 4 is centered in a narrow band near 1.38 microns. We can tell that the band is narrow because the spectral response function is not very wide. At this wavelength, the transmissivity of the atmosphere is near zero. This is because water vapor absorbs radiation at this wavelength. Obviously then, the purpose of this band is not to detect radiation scattered off the ground or ocean. Instead, its primary purpose is to detect high cirrus cloud. There are many scenes that—using visible imagery—show a variety of cloud at various altitudes. Sometimes, however, cirrus cloud is so physically and optically thin that it is difficult to see in visible imagery. However, by choosing to passively observe radiation scattered in a band subject to absorption by water vapor, we only end up seeing radiation that is scattered by clouds at the top of the atmosphere. We don't see clouds lower down because the 1.4-micron radiation is absorbed and re-emitted at a much longer wavelength consistent with its temperature.

Slide 8: For example, let's look at a sample of GOES-17 imagery over the default 1-minute mesoscale sector located over California. First, we'll look at a morning scene in Band 2, the highest-resolution red band. What is plotted is the **brightness value**, which is essentially just normalized radiance as we will see in the first lab. Intuitively, the white, or bright areas, indicate where more solar radiation is reflected. An expansive offshore deck of stratocumulus cloud is visible, as well as some reflection off the surface over the Central Valley or in Nevada. There are also obviously some high clouds visible extending from southwest to northeast in the image. With the morning insolation angle, the higher clouds stand out easily from the background stratocumulus deck because of the shadows they cast on the lower layer.

Slide 9: If we look at reflection in Channel 4 at nearly the same time. Reflection of the low-level cloud or surface are no longer visible. You only see reflection off of the high cloud because radiation that passes further down into the troposphere is absorbed by water vapor and so cannot be reflected to the satellite sensor. However, radiation reflected off of clouds high in the troposphere do not encounter as much water vapor and is more likely to reach the satellite sensor. As you can see, this channel is useful for quickly separating cirrus cloud from other reflective clouds or surfaces below. It could be useful for quickly identifying the direction of upper-level flow in a complex scene with many clouds moving in various directions in a directionally sheared environment.

Slide 10: Recall the general equation for radiative transfer in a scattering atmosphere. This would apply primarily to all but Channel 4, for which we would have to consider absorption and scattering. L_0 in the equation was the surface radiance directed upward and caused by reflection of radiation. However, how radiation is reflected off the surface is dependent upon many properties of the surface and the angle at which radiation is incident upon the surface. The bidirectional reflectance is a function that describes the scatter of radiation coming from a direction theta_r and phi_r into many directions theta_i and phi_i. We won't go into detail about what these functions look like other than to conceptually describe two types of scatter off the surface. One is Lambertian reflectance. It occurs when incident radiation is scattered isotropically, or equally in all directions, off of a surface. The other is specular reflection, or mirror-like reflection in which most radiation scattered is focused onto one angle. In the real world, perfectly isotropic or specular reflection rarely occurs off the surface, but there are situations in which the ocean may behave in one of these extremes. A very rough ocean may behave more like an isotropic scatterer, especially if the incident radiation is from above. A very calm ocean surface can act like a mirror and focus radiation along a narrow return angle. This appears to the satellite as a locally heightened area of reflectance, known as sunglint.

xxxSlide 11: An example of sunglint can be seen from some old GOES-10 imagery from the previous century. Although faint, inside the orange circle, you can see how the reflectance off of the ocean surface looks locally higher than in nearby locations.

Slide 12: A clearer example can be seen in more recent imagery from GOES-16. Again, reflectance is enhanced in the orange circle. You could especially tell that this is caused by sunglint because if you view a loop of shortwave imagery, you will see the area of heightened reflectance move westward during the day as the Earth rotates.

Slide 13: The state of the sea surface affects the appearance of sunglint. Shown here is imagery from the Advanced Very High Resolution Radiometer, which was once actually very high resolution for our capabilities at the time the instrument was developed. The region of sunglint off the west coast of Mexico is apparent.

Slide 14: Compare that sunglint with that from the same location at nearly the same time a few days later. The solar angle is essentially the same, but the sunglint reflectance is much greater and more focused in a narrow region off the coast of Mexico. This heightened sunglint can occur if the sea state is calm and the ocean behaves more like a specular reflector. The variability of the bidirectional reflectance of the ocean surface will again be considered later in the course when we discuss the use of scatterometers for estimation of surface wind vectors.

Slide 15: Let’s take a look at a couple of other bands. In particular, we’re look at Bands 1 and 3.

Slide 16: Here’s another look at Band 2 from our scene over California.

Slide 17: And here is the same scene, a few minutes later, as imaged in Band 1. If we flip back and forth, you can see that radiation in Band 1 is generally reflected more than that in Band 2. This is because blue light is more likely to be scattered by air molecules and aerosols in the atmosphere than red light. Therefore, Band 1 is often used for aerosol detection, such as dust, haze, or smoke.

Slides 18 and 19: Channel 3 detects near-infrared radiation. It is not visible to the human eye and is a little longer in wavelength than red light. Vegetation tends to absorb red visible light and reflect the near IR radiation. Therefore, land surfaces often show up as highly reflective. Notice how in this scene, the Central Valley is much more reflective in Band 3 than it was in

Bands 1 or 2. In fact, much of California is more reflective in Band 3. Take a look at the "true color" composite image of the same scene, and flip back and forth. The highest reflectances in Band 3 are where green colors—or vegetation—appear. In the Central Valley, this Is active agriculture. In northern California along the coast and through the Sierra, forests reflect near IR radiation efficiently. In contrast, plants absorb more visible light, which is why the ground is not highly reflective.

Slide 20: We won't spend much time in this class looking at data from Bands 5 and 6; however, they also have specific useful applications. Band 5 is centered at 1.61 μm; at this wavelength, ice absorbs radiation more than liquid water. This means that ice clouds will appear to be less reflective than lower liquid water clouds, which can help infer the phase of water in the clouds. If you compare this image to the other images from this scene, you'll see that the low-level stratocumulus cloud appears relatively brighter compared to its surroundings than in other channels. Shadows of high cloud also are contrasted against the low cloud background.

Slide 21: Channel 6 is used for cloud particle size detection, which is an application we will not cover in this class. However, it can also be used to detect hot fires at night. Since no solar reflectance will appear at nighttime, hot fires that emit partially in shortwave wavelengths appear in this channel and are strongly contrasted against the relatively dark background at night. In the next module, we will look through the longwave GOES channels that detect radiation emitted from Earth.

Module 2.3

Slide 1: We continue our discussion of GOES imagery and data in this module, which focuses on the longwave bands.

Slide 2: Remember the second idealized case covered in a previous module. It related to passive remote sensing of IR radiation emitted by both the surface and along a path through the atmosphere to a satellite. We ended up with two terms to describe the radiance at the top of the atmosphere. The first described the percentage of radiation emitted by the surface that was not absorbed by the atmosphere, and the second quantified the total radiation emitted along the path at the wavelength of interest, which is a function of the atmospheric temperature and optical depth profiles. Ultimately, emissions observed at a satellite are a summed combination of surface emissions, if any, and emissions occurring at various temperatures throughout the depth of the atmosphere. We are not concerned about scattering of IR by air molecules in a clear-sky environment; however, clouds, especially physically thick ones, do scatter some IR radiation. Liquid water and ice also absorb some IR radiation.

Slide 3: Let’s return to our plots of spectral response functions. In this module, we turn our attention to the bottom row and focus on the SRFs for IR bands on GOES. The bands are generally wider than for the shortwave bands; this is required since in general, longwave spectral radiance is less than that at shorter wavelengths; therefore, wider bands are required to achieve the desired spatial resolution. The red line denotes the average brightness temperature as a function of wavelength given standard vertical profiles of temperature and humidity. Higher infrared brightness temperatures correspond with higher atmospheric transmissivity.

Slide 4: The brightness temperature is the temperature that a blackbody would require to emit the radiance observed by a satellite. Consider the expression for Planck blackbody radiance, which is a function of wavelength and temperature. By rearranging the equation, we can easily derive an expression for brightness temperature that is a function of top of atmosphere radiance observed by a satellite at some wavelength. An example of brightness temperatures for a band in the IR atmospheric window is shown here. This is an example of an enhanced image; the colors indicate objects—often clouds—that have the lowest brightness temperatures. Darker colors indicate warmer emissions. Since temperature generally decreases with height in the troposphere, the darker regions correspond to emissions from lower in the atmosphere or even the surface. An unenhanced image would simply replace the colors with various shades of gray and white, with the brightest whites denoting the lowest brightness temperatures.

Slide 5: A couple of things must be considered when interpreting brightness temperatures. First, the emitting body might not be a blackbody; however, for infrared radiation, we will generally approximate the surface and clouds as blackbodies for the purposes of discussion in this class, meaning that epsilon can be approximated as one. This means that a lower IR brightness temperature generally corresponds to a lower actual temperature. When we discuss microwave radiation later in the class, we will see that surfaces can emit much less than blackbodies, which greatly alters interpretation of brightness temperature. Second, extinction of radiation could occur as it propagates upward through the atmosphere. Furthermore, the rate of extinction may change with height. Therefore, the brightness temperature observed may be representative of radiation emitted from various different altitudes, including the surface. The vertical derivative of direct transmittance is known as the weighting function, which will be covered with more detail in the next module.

Slide 6: Now, returning to our plot of SRFs, we see that some bands—like 8, 9, and 10—sit in parts of the spectrum where average brightness temperature is low. The brightness temperature is low between roughly 5 and 8 µm because water vapor absorbs radiation. Thus, much of the radiation emitted by the surface at these wavelengths gets absorbed as it passes upward through the atmosphere. Whatever object that absorbs the radiation then re-emits radiation, but it preferentially does so at a wavelength consistent with its temperature—usually a slightly longer wavelength since temperature generally decreases with height in the troposphere. Next, let's step through each band individually.

Slide 7: The shortwave infrared band is the only GOES band that routinely observes overlap between reflected solar and emitted terrestrial radiation. Generally, the radiance observed is converted to brightness temperature for display. This daytime image in Band 7 displays high clouds as black. While they may be cold, they are also reflective of solar radiation. The shortwave IR is particularly useful for fire identification similarly to Band 6. It is also useful for detecting small ice crystals because they are more reflective of radiation at this wavelength. This increases the radiance at the sensor, making the cloud appear "warm". This is why the high clouds appear as black in this image. Larger ice crystals are not efficient at scattering 3.9 µm radiation; therefore, they appear as cold.

Slide 8: Bands 8 through 10 are the "water vapor bands". As stated before, the bands are located in a region of the EM spectrum where water vapor absorbs. That means that the horizontal variability in brightness temperature observed in these bands is caused by lateral variability in the vertical distribution of water vapor.

Slide 9: These channels are best interpreted as a group. Using the same color scale, we see many more red and orange colors at Band 9 except for where high clouds are present.

Slide 10: The brightness temperature is even warmer for Channel 10, with brightness temperature much closer to 0°C, as denoted by the red shading.

Slide 11: When we compare the three alongside each other, when using the same color scale, it is clear that the average brightness temperature of the same scene generally increases moving from Band 8 to Band 10. This is consistent with the graph in the bottom left, in which brightness temperature increases going down on the y-axis, and is therefore greater in Band 10 than in Bands 9 and 8. This means that of the three, Band 8 is most susceptible to absorption by water vapor in the atmosphere. Emissions from the surface at 6 µm wavelength are less likely to penetrate to space than 7 µm radiation. Therefore, in Band 8, we see more emissions from high in the atmosphere, and thus the brightness temperature is lower. In very dry areas, the difference in brightness temperature will be less between the three bands. It is not unusual for a particularly dry column to yield a brightness temperature of 0°C at Band 8. The brightness temperature of the three bands in a region of thick clouds will be very similar because few emissions from beneath cloud top reach space at any wavelength in this part of the IR spectrum.

Slide 12: We will not discuss Channel 11 all that much, but the 8.5 µm band is situated partially in an absorption band for sulfur dioxide, which is emitted by volcanic eruptions, making this channel useful for detecting volcanic ash. Otherwise, the atmosphere is mostly transparent to this radiation. Also, the emissivity of liquid water is less at 8.5 µm than at other bands in the atmospheric window; therefore, this band offers another way to detect the phase of water in clouds.

Slide 13: Another band, Channel 12, sits in the middle of the ozone absorption band. You can immediately see how the brightness temperature in Band 12 is generally lower than other bands. This is because ozone emissions in the lower stratosphere occur at a level where the temperature is colder than much of the troposphere. However, in deep convection, Band 12 brightness temperature can be higher than in an atmospheric window channel that is not sensitive to ozone because the lower stratosphere is warmer than the upper troposphere.

Slide 14: Channel 13 at 10.3 µm is the least impacted by absorption by atmospheric water vapor. Therefore, it is most commonly used for 24-hour surveillance of clouds. Remember, the brightness temperature is not necessarily temperature. The high cloud seen over California and offshore is obvious in this imagery. The brightness temperature of the cloud ranges from roughly -10°C to -30°C. However, some of the cloud may actually be much colder than indicated by the brightness temperature, especially if it is physically—and therefore optically—thin. Thin cirrus clouds often allow surface radiation to penetrate through them, meaning that radiance detected at the satellite sensor is caused by some combination of emissions at the top of the cirrus cloud and emissions from below, which is exactly what Schwarzchild's equation says we should expect.

Slide 15: Channel 14 is similar to 13 but experiences a little more absorption by water vapor at 11.2 µm. It can generally be used for the same purposes as Channel 13, but the two together can be used to assess atmospheric water vapor concentrations.

Slide 16: Channel 15 at 12.3 µm also sits in the atmospheric window but experiences a little more absorption by water vapor. The split window difference is calculated using Channels 13 and 15; it is just the difference in brightness temperature between the 12.3 and 10.3 µm brightness temperatures. This difference increases as atmospheric water vapor increases. Emissivity of dust and ash is lower at 10.3 µm, so brightness temperatures will be larger in Channel 15 than in Channel 13 in areas where such aerosols are present, potentially making the split window difference negative.

Slide 17: Band 16 will not be used much in this class either. It sits on the edge of a broad carbon dioxide absorption band. Brightness temperatures will generally be cooler in Channel 16 than in the center of the atmospheric window, where the SRFs for Channels 13 through 15 are situated. Brightness temperature from this band is also used to generate many post-processed satellite-derived products.

Module 2.4

Slide 1: In this module, I'll briefly introduce the concept of weighting functions and show a couple of examples. Weighting functions are useful for showing from where radiation in a column that reaches a satellite is emitted. In this class, we will see weighting functions for passive sensing applications in the infrared and microwave parts of the EM spectrum.

xxxSlide 2: Recall the definition of direct transmittance; it is the exponential of the negative vertical path optical depth of the atmosphere above a point normalized by the exitance angle to correct for the path length. If we take the derivative of direct transmittance with respect to pressure, we get an expression that contains an additional coefficient of 1 over mu and the vertical derivative of optical depth. If we take the below expression for Schwarzchild's equation, we can replace the exponential in the second term with the weighting function *W*. The weighting function tells us the relative contributions of emissions for various pressure levels to the top of atmosphere radiance. Let's take a look at a few simple examples.

Slide 3: Suppose an atmosphere has an optical depth profile that looks that the plot in the middle. This is pretty typical in wavelengths that are susceptible to absorption by water vapor, whose concentration roughly decays exponentially with height to first order. The direct transmittance of the atmosphere for the given optical depth is shown at left and decays to zero near the surface. The weighting function is shown at far right and is the vertical derivative, in pressure coordinates, of the direct transmittance. Thus, it will peak in magnitude where the direct transparency changes the most. This happens at around 2 km altitude in this example. The weighting function tells us that radiation that would reach the satellite comes from mostly between the surface and 5 km, although a little radiation comes from even higher altitudes since the direct transmittance does not go quickly to 1 at high altitude.

Slide 4: In the second example, we stretch out the vertical profile of optical depth. It begins to increase (going from top down) at around 10 km, and a little below 10 km is where the direct transmittance begins to decay quickly. Therefore, the weighting function peaks around 8 km; however, it is nonzero through the depth of the troposphere. Most of the radiation reaching space, however, comes from roughly between 5 and 11 km.

xxxSlide 5: One last example shows a weighting function with an upper tropospheric maximum. This might represent a wavelength that is sensitive to something present high in the atmosphere, such as molecular oxygen or ozone. In this case, very little of the top of atmosphere radiance is influenced by emissions from the lower troposphere because any emissions are absorbed and re-emitted higher in the atmosphere.

Slide 6: For bands that are sensitive to molecules that vary in concentration horizontally, like water vapor, the weighting function will vary substantially as well. Suppose you are observing the same scene from a satellite at two different times. One time is moister than the other and has a mixing ratio profile that looks something like the green line. Later, the atmosphere is much drier, and the mixing ratio profile more closely resembles the red line. In a water vapor absorption band, the volume absorption coefficient would vary with the mixing ratio, so the optical depth profile has a similar shape to the mixing ratio profile. We can then analytically compute the direct transmittance and weighting function. Both tell you that top-of-atmosphere emissions generally come from lower altitudes in the dry atmosphere than in the moist one, although some overlap exists between the two weighting functions. Because the transmissivity of the moist atmosphere goes to zero above the ground, the weighting function in the moist atmosphere is zero below that point. Finally, the bottom left plot shows how direct transmittance might vary as a function of wavelength in a band centered around the black line that is located somewhere in a water vapor absorption band. Direct transmittance in the moister environment will be small—perhaps zero—over a wider band surrounding the central wavelength than the dry environment, in which the direct transmittance might not even reach completely zero at the surface.

Slide 7: Weighting functions also differ between wavelength bands. Recall the SRFs for the GOES longwave bands. Channels 8 through 10 are within a water vapor absorption band, but Band 8 experiences more absorption than Band 10. Therefore, in the same environment, the weighting function for Band 10 will tend to have a peak at a lower altitude than Band 8—consistent with the notion that the typical brightness temperature for Band 10 is higher.

Slide 8: We will wrap up by taking a look at two different sets of weighting functions. The first is at Shelby, Alabama. A sounding launch indicated almost 50 mm of total precipitable water, which is indicative of a moist mid-latitude environment. The temperature and dewpoint are respectively shown as solid and dashed lines on the right panel, and the two are near each other, especially in the upper troposphere. The green, blue, and purple lines denote the weighting functions for GOES Bands 8, 9, and 10 at this location. Band 8 has a maximum near 200 mb, while Band 10 has a maximum near 400 mb. However, Band 10 detects some radiation from as low as 700 mb. Band 8 detects no radiation below about 400 mb. The weighting function for Band 9 sits between the other two.

Slide 9: This figure shows the weighting function at Las Vegas, Nevada, at the same time. Las Vegas is in a desert and this sounding was taken in July. The total precipitable water was a fairly dry 13 mm. When compared to the weighting functions in the moister Alabama environment, these three weighting functions—still representing Bands 8 through 10—are much closer together. They all have peaks between about 450 and 550 mb. However, Band 10 still sees radiation from farther down in the troposphere than Bands 8 or 9; therefore, we should expect that the Band 10 brightness temperature will be the warmest of the three at this location. The weighting functions shown in this and the previous slide are based on temperature and humidity observations from weather balloons. They give us a convenient way of quickly interpreting the radiances viewed by satellites. We will return occasionally to the topic of weighting functions as we continue to discuss passive remote sensing.

Module 2.5

Slide 1: This short module will introduce you to the Joint Polar Satellite System (JPSS), a cooperative Japanese-US effort.

Slide 2: Currently, JPSS consists of two near-polar orbiting satellites in the same sun-synchronous orbit: Suomi-NPP, the Japanese satellite, and NOAA-20, the U.S. satellite. Three satellites are planned for launch over the next decade. The satellites are in daytime ascending orbits with equator crossing times of about 1:30PM. NOAA-20 lags Suomi-NPP by about 50 minutes such that the sub-satellite nadir point of NOAA-20 is a little to the west of that of Suomi-NPP. The satellites provide high-quality data in locations where GOES cannot, such as in high latitudes, where the poles are observed once per hour. The addition of future satellites will increase the temporal resolution at the poles. Data with slightly higher resolution than GOES—especially in the IR channels—is also collected in the middle and low latitudes, although unlike GOES, the data is collected in swaths about 3,000 km wide.

Slide 3: The figure here shows how the two orbits of Suomi-NPP and NOAA-20 relate to each other. As NOAA-20 ascends, Suomi-NPP descends on the opposite side of Earth. The satellites have nighttime equator crossing times of about 1:30AM such that the two platforms cross the equator moving in different directions on opposite sides of Earth.

Slide 4: NOAA-20 carries five main instruments. The Cross-track Infrared Sounder, or CrIS, is a sounder with high spectral resolution, detecting radiation in over 3,000 spectral bands. The locations of the bands are on the edges of water vapor and oxygen absorption bands, meaning that the weighting functions of the various channels peak at different levels in the atmosphere. This makes CrIS a suitable instrument for estimating vertical profiles of temperature and humidity. The Advanced Technology Microwave Sounder, or ATMS, has 22 channels in the microwave part of the spectrum. It generally has lower spatial resolution than CrIS; however, because microwave radiation is not as efficiently scattered by clouds as infrared radiation, ATMS can collect vertical profiles of temperature and humidity in cloudy areas, and together with CrIS, provide comprehensive coverage of the entire planet.

Slide 5: The CERES Flight Model 6 instrument is the latest in a series of CERES instruments dating back to 1997. It is capable of providing accurate measurements of surface and top of atmosphere radiative fluxes. Based on observed clouds, it also provides estimates of coarse atmospheric profiles of radiative fluxes that are consistent with derived surface and observed top of atmosphere fluxes. This provides a useful tool for a variety of research applications.

Slide 6: The Ozone Mapping and Profiler Suite (OMPS) maps column-integrated ozone in 50 km wide boxes. Profiles of ozone concentration are also estimated but have lower spatial resolution. These are useful for tracking tropospheric ozone and monitoring stratospheric ozone at the poles. The Visible Infrared and Imaging Radiometer (VIIRS) is most like the ABI on GOES. It passively detects radiation from a variety of shortwave and longwave bands in the visible, near IR, and terrestrial IR.

Slide 7: A summary of the VIIRS bands is seen here. Note that multiple visible bands are used, contrasted with just the 2 or 3 on GOES or Himawari. Many bands are clustered in the violet to blue part of the visible spectrum. These are useful for deriving ocean color in combination with the green and red visible bands, including quantities such as chlorophyll concentration, total absorption, or particulate backscatter. VIIRS bands are also used for detection of aerosols, which more easily scatter low wavelength light. Quantities such as aerosol optical depth can be derived from VIIRS visible reflectances. The low Earth orbit of the JPSS satellites also allows infrared data collected to have higher spatial resolution. At nadir, resolution is nominally 742 by 776 meters or better in all channels. VIIRS can also provide estimates of soil moisture and ice surface temperature and is useful for monitoring of sea ice.

Slide 8: An example of a daily composite of VIIRS derived RGB imagery from NOAA-20 is shown here. You can see how far successive orbits are because of swaths of sun glint that appear equally spaced apart.

Slide 9: Monthly averaged global chlorophyll concentrations derived from Suomi-NPP VIIRS are seen here. This product is available at 4 km grid spacing. The warm colors represent enhanced chlorophyll concentrations, which is indicative of active phytoplankton. Several dead zones can be seen in purple and blue.

Slide 10: We can zoom in on various parts of the world to see various quantities in greater detail. For example, enough spatial resolution is available to see differences in chlorophyll concentrations on opposite sides of Monterey Bay.

Slide 11: An example of aerosol optical depth in one visible band is shown here. Some features stand out in this snapshot. Saharan dust is denoted by a bright red area over Western Africa. Aerosols in northeastern China are seen, and pollution in northern India is visible. Biomass burning in equatorial Africa is also apparent. Click the link in the slides to see an animation of this product over a three-month long period in 2020 at the beginning of the COVID-19 pandemic. The JPSS satellites captured some regional variability in aerosols in normally heavily polluted areas such as northern India, although you'll notice that biomass burning persists in many places, especially in south Asia.

Slide 12: Brightness temperatures at various bands can be displayed for the CrIS instrument on NOAA-20; shown here is one example at 11 microns. It is obvious from these two panels that NOAA-20 is in a daytime ascending orbit.

Slide 13: For comparison, ATMS brightness temperature is shown at 23.8 GHz. We will discuss microwave data more later in the quarter, but you may note how brightness temperatures are markedly different from IR brightness temperatures. A swath of heightened microwave brightness temperature is present in each swath. It is the manifestation of sun glint of solar microwave radiation reflected off of Earth.

Slide 14: Finally, the total column-integrated ozone product is shown from Suomi-NPP. Enhanced ozone is seen in by the red and yellow colors, while depleted ozone is obvious near the South Pole in this image.

Module 2.6

Slide 1: This module will show some examples of how infrared radiances are used to estimate sea surface temperature or land temperature.

Slide 2: As we recall from previous modules, brightness temperature is not the actual temperature of an object. The brightness temperature in a cloud-free environment is not the surface temperature because 1) The atmosphere above may absorb small amounts of radiation emitted by the surface, even in atmospheric windows that are not perfectly transparent, or 2) The surface is not a blackbody. The brightness temperature is calculated as a function of satellite-observed radiance. The goal for estimating SST is to calculate a brightness temperature given an observed radiance and correcting that brightness temperature using radiances in multiple bands to correct for water vapor absorption. Of course, in cloud free areas, surface emissions are not clearly observed.

Slide 3: In Schwarzchild's equation, we seek to estimate SST using radiances in bands in which the first term on the right-hand side of the displayed equation dominates the second term. This means that the direct transmittance, tau, needs to be large, or that the weighting function needs to have as large a value as possible at the surface.

Slide 4: This occurs in atmospheric window channels. The plot shows direct transmissivity of the atmosphere as a function of wavelength and the contribution of each of several atmospheric constituents to total extinction of radiation along a vertical path through that atmosphere. As you know by now, total extinction is limited at shortwave IR wavelengths around 4 microns and between 8 and 12 microns with the exception of the ozone band. Even within the IR atmospheric window, different bands experience slightly different sensitivity to water vapor, and the small differences in brightness temperature in such bands can help quantify the column-integrated water vapor present. For example, AVHRR, the visible and IR satellite sensor on NOAA-19, uses a shortwave IR channel and two terrestrial IR channels to estimate SST. Differences between AVHRR Channels 4 and 5, the red and blue shading, will occur because of variations in column water vapor because other molecules have negligible effect on absorption at 10–12 microns.

Slide 5: The GOES channels used to compute SST are highlighted here: Three in the terrestrial atmospheric window, a shortwave IR band, and one on the shoulder of the water vapor absorption band.

Slide 6: Let's think about what an increase in water vapor then does to the brightness temperature. Suppose we have a homogenous single-layer atmosphere with direct transmissivity of 90% and some temperature, and a blackbody surface with a higher temperature. We can simplify the integral since T is constant with height. By plugging in 0.9 for tau, then the top of atmosphere radiance is a weighted average of surface emissions and a small amount from the atmosphere. Therefore, the brightness temperature is not exactly the surface temperature but is instead slightly lower.

Slide 7: Observing this simple atmosphere with two channels allows us to incorporate the derived water vapor path length into a corrected temperature that more closely resembles the true blackbody emission temperature of the surface. As we'll see here, higher concentrations of water vapor increase the difference in brightness temperatures between bands that have slightly different sensitivities to water vapor absorption. Suppose that the same atmosphere has a transmissivity of 95% to 11-micron radiation and 90% to 12-micron radiation, which are approximately the center wavelengths of AVHRR Channels 4 and 5. We can solve for the optical depths of the atmosphere at each wavelength using the relationship between optical depth and direct transmittance. In this simple example, we get optical thicknesses of about 0.051 and 0.105 for the two channels. The difference in radiances is about 5% of the difference between brightness temperatures. However, if we double water vapor concentration and carry out the same calculations, we find that the difference in radiances between the two bands goes up only to 9% of the difference in brightness temperatures. The point here is that water vapor path in the atmosphere does not correspond linearly with changes to split window differences between neighboring IR bands. Ultimately, we require somewhat complicated empirically derived relationships to estimate SST based on brightness temperatures.

Slide 8: The AVHRR multi-channel sea surface temperature estimate derives surface temperature as a linear combination of each channel's brightness temperature plus the difference in brightness temperature. The relationship between brightness temperatures and SST therefore takes on some linear form such that SST equals a constant plus unknown coefficients times brightness temperatures for various bands. The coefficients are empirically determined by comparing known observations of SST, such as through ship or buoy data, to satellite brightness temperatures. At night, we can use the shortwave IR wavelength, but we cannot do so during the day because of additional reflected solar radiation.

Slide 9: The following shows some different versions of the multi-channel SST relationships used without the coefficients filled in. Note that the nighttime relationships incorporate three bands, one of which is the shortwave IR. All relationships include a term that corrects for off-nadir observations along increased path lengths that increase the amount of absorption that occurs. Different satellites—GOES or the JPSS satellites—use different coefficients given the available bands on each platform.

Slide 10: Consider that we are only detecting information about the very top of the ocean surface. Liquid water scatters IR radiation efficiently; therefore, emissions by water molecules below the upper 1 millimeter or so are absorbed before they escape the water. Therefore, SST estimates made with infrared are "skin" temperature estimates, capturing the temperature of just a thin layer. Microwave-derived brightness temperatures represent SST over a slightly deeper layer because microwaves are not as susceptible to scattering by liquid water.

Slide 11: A sea surface temperature product can then be produced that covers much of the globe. This example is derived from multiple sources, including buoy data, but we will see in the lab that satellite brightness temperatures can be reliably converted to SST with some limitations. In this plot, the warmest sea surface temperatures are found over the tropical warm pool over the West Pacific. Brightness temperatures in clear air here would therefore be quite warm; although, large differences in atmospheric window channels might exist because column water vapor is often high.

Slide 12: The SST data can also be used to produce products such as the weekly SST anomaly seen here. It shows a prominent positive SST anomaly in the North Pacific near the Gulf of Alaska and perhaps a weak La Niña temperature distribution in the equatorial East Pacific.

Slide 13: Similar procedures can be used to estimate land surface temperature or moisture. We won't explore these in detail in this module, but we will briefly consider that objects over land are diverse, and many are not even close to approximate blackbodies.

Slide 14: An example of emissivity for a few various surfaces is shown at bottom, with the direct transmittance of the atmosphere shown at top. Some of the surfaces plotted are close to perfect emitters in the IR. However, others are not. For example, seawater, although a good emitter in the atmospheric window, is actually not a good emitter at 4 microns. Sandy soil cannot be approximated as a blackbody between roughly 3 and 5 microns and 8 to 11 microns. This means that brightness temperatures over such surfaces at the wavelengths listed will be low—not necessarily because the atmosphere is opaque—but because the emissivity, and therefore the first term on the right-hand side of the Schwarzchild's equation as shown earlier in this module, is small relative to what it would be if we assumed epsilon equals 1.

Slide 15: For the most part, vegetation is an efficient emitter at IR wavelengths, although dry grass is shown here to have emissivity of only 90% between 4 and 5.5 microns.

Slide 16: Man-made surfaces also have variable emissivity. Asphalt, concrete, and steel are examples of objects with high emissivity at a variety of IR wavelengths. However, aluminum and brass are extremely poor emitters. Even if hot, emitted radiance, and thus brightness temperature, would be quite low.

Module 2.7

Slide 1: This module will explore a few unclassified sources of high-resolution satellite imagery, and some of its utility. We will also cover a couple of different vegetation indices that can be derived from this data.

Slide 2: We'll start off with a panchromatic image of San Francisco, from a European satellite, SPOT-5, taken in 2002. We'll define panchromatic in a few slides. The imagery was collected with spatial resolution of 5 meters by 5 meters, meaning that each pixel of data represents a 25 square meter area on the surface. Collection of high-resolution data is possible by using sensitive sensors and large mirrors in low-Earth orbit. We'll discuss a couple of satellite platforms in sun-synchronous orbit that deliver data that is capable of making meaningful observation of the land surface with detail: Landsat and WorldView.

Slide 3: Landsat-8 is the most recent in the Landsat series, which dates back to 1972. The satellites are publicly funded and data is provided by the US Geological Survey. It consists of two instruments: The Operational Land Imager, which is responsible for passive detection of visible and near-IR reflectance; and the Thermal Infrared Sensor, which detects IR bands in the atmospheric window. OLI channels have spatial resolution of 30 by 30 meters, or 15 by 15 meters in the panchromatic band. TIRS data has spatial resolution of 100 by 100 meters. The orbit is daytime descending with an equatorial crossing time shortly after 10AM.

Slide 4: Shown are the bands available from Landsat-7, which is also in operation. It consists of three visible bands, a near IR band, two shortwave IR bands, one thermal infrared band, and a panchromatic shortwave band.

Slide 5: Landsat-8 adds on three bands: An additional visible band in the blue-violet for enhanced aerosol detection, a cirrus band (similar to that available on GOES), and a second thermal infrared band, which combined with the other is useful for deriving soil moisture from IR radiances.

Slide 6: What we have seen in previous modules is known as multispectral imagery—essentially just various channels in different bands. True color imagery derived from GOES, for example, is derived from a multispectral combination of bands in the shortwave. Panchromatic bands are a single wide band in the visible spectrum. They are usually the highest-resolution channel available on an Earth-observation satellite; however, panchromatic data will often be presented in grayscale. The act of pan-sharpening is the act of using the red, green, blue, and panchromatic channels together to approximately produce an accurate representation of true surface color at the high resolution of the panchromatic band.

Slide 7: Spectral response functions for the four narrow visible bands and the panchromatic band on Landsat-8 are shown here. Note how the SRFs for Channels 1 through 4 are separated from each other, capturing reflected radiation in different colors. The panchromatic band, however, spans many different colors, running from 503 to 676 nm, and including the entirety of the red and green bands, and a small part of the blue band. Therefore, true color imagery cannot be derived from just the panchromatic band; however, because the band is very wide, it senses more radiation than any of the narrower channels. As a result, it can detect enough signal to capture reflectance at smaller spatial intervals.

Slide 8: WorldView satellites are another series that has produced publicly available data. However, like many other private satellite systems with high-resolution data, the data is proprietary. The U.S. government has purchased much of this data, and we will see an example shortly. WorldView is in a similar orbit as Landsat-8, although about 100 km lower. It samples about 680,000 square kilometers daily. In that area, data is collected in a variety of spectral bands, seen in the leftmost part of the table. The panchromatic band has spatial resolution of approximately 30 by 30 *centi*meters, meaning that it can resolve small details on the surface. For example, if a typical human laid on the ground, they would take up about 3 data points in WorldView. The maximum allowable resolution of a private U.S. satellite is 25 by 25 centimeters. Classified government satellites are known to use large mirrors (comparable to that used on space telescopes) that permit even higher resolution. The satellite from which a certain U.S. government civilian leaked imagery on Twitter is thought to have spatial resolution of approximately 10 by 10 centimeters.

Slide 9: For these super high-resolution satellites, we are mostly concerned with viewing the surface, although recent research, including some of my own, seeks to leverage public high-resolution data to study tropical clouds. Consider the Schwarzchild's equation for scattering at top. We want the second term to be minimized as much as possible to gain maximum detail about the surface. The biggest challenge is scatter by clouds. This should be rather intuitive. If a cloud is between the satellite and the ground, the satellite can't see reflected radiation from the ground. This problem can't be overcome; simply an end-user must wait for a subsequent satellite pass without cloud to see the surface at that location. However, aerosols and air molecules also cause scatter. However, both of these effects can be estimated to correct observed reflectances to a value more representative of the true surface properties. For emitted IR channels, as with previous applications such as SST estimation, we want to maximize the first term of the Schwarzchild's formulation shown at bottom. Ideally, optical depth is small and emissions are representative of the surface. If so, the two terrestrial IR radiances from Landsat-8 can be compared to estimate soil moisture, which is useful for drought monitoring.

Slide 10: True color imagery from Landsat-8 over Fiery Cross Reef in the South China Sea is shown here. In this well-known example, we can see the progression of a man-made island built in the South China Sea in a location where higher reflectance from the ocean—hinting at a shallower depth—was located.

Slide 11: In comparison, look at the resolution achievable by WorldView satellites with a 30-centimeter panchromatic band. This image was collected from an orbital altitude of over 600 km.

Slide 12: Another example of Landsat-8 RGB-derived imagery is shown for an entire “tile” of data—this one over the Santa Lucia Mountains along the California Central Coast in 2019.

Slide 13: Previously, we considered that different types of surfaces have different emissivity coefficients. Likewise, the reflectance of the surface varies based on the composition of the surface. This plot depicts reflectance as a function of shortwave wavelength for a variety of different surfaces. Note how plants—the pink, yellow, and light blue lines denoting grass and deciduous and conifer trees—are not very reflective of visible light, although, a small peak in reflectance is seen around 550 nm. You should be able to deduce what color that corresponds to. Just beyond the red end of the visible spectrum, reflectances of plants increase to about 50%. We saw this in the first lab as well; Channel 3 on GOES was much more reflective in vegetated areas than Channels 1 or 2. This is because plants reflect near-infrared radiation.

Slide 14: This fact can be used to compare reflectances in the visible and near IR wavelengths to deduce vegetation on the surface. One such index to do so is called the normalized difference vegetation index, or NDVI. It is simply the quotient of the difference between near-IR and red reflectances and the sum of the two. Green forests would reflect lots of near-IR but make the visible reflectance small because leaves absorb red light for photosynthesis. Thus, the numerator would be large. NDVI ranges from –1 to 1. A value of over 0.5 usually indicates lush vegetation, 0.2–0.5 often represents sparse vegetation, and under 0.2 includes non-vegetated areas such as water, deserts, ice, or man-made structures.

Slide 15: Many other indices exist for estimating vegetation robustness. An alternative to the NDVI is the Enhanced Vegetation Index, or EVI, which incorporates multiple visible bands and is more sensitive to vegetative properties such as leaf area index.

Slide 16: For the same Landsat-8 image of the Santa Lucia mountains shown earlier, the NDVI is shown. Dark green areas indicate high NDVI. These are present in vegetated areas; if you zoom in on your slides for this module, you can see some individual active agricultural plots. Forests along the Big Sur coastline also show enhanced NDVI. Urban areas, such as the Highway 101 corridor and some small cities, are represented by brown and white colors. Yellow-green colors are representative of areas that are either sparsely vegetated, or consist of dry, brown grass that does not absorb much red light. These types of indices can be computed using a variety of different datasets that collect information in the appropriate bands and can be particularly useful, as an example, for agricultural monitoring over large areas.