Module 4.1

Slide 1: This module begins a series of videos that will explore remote sensing in the microwave portion of the EM spectrum. In this module, we will look into some of the basic similarities and differences between passive remote sensing in the IR or visible wavelengths compared to microwave frequencies.

Slide 2: Recall from the beginning of the course that the microwave part of the EM spectrum encompasses much longer wavelengths than IR or visible radiation. Microwave radiation that we will discuss in this class is situated mostly in a region of medium to high atmospheric absorptivity, so attenuation or extinction of radiation is a big concern for many microwave frequencies.

Slide 3: Because of its longer wavelength, microwave radiation is generally not as susceptible to scattering as shorter IR and visible radiation. Therefore, microwaves can more efficiently penetrate features that IR and visible light cannot. In particular, microwaves are used to see emissions through clouds and layers of liquid water and soil. However, emittance of microwave radiation by Earth is much smaller than its infrared emissions. Therefore, the spatial resolution of microwave data is generally less than that of IR data. This also requires that passive and active microwave instruments operate in low Earth orbit. As a result, the temporal resolution of space-based microwave data—also the return period of a satellite to a point on the surface—can be several days. However, constellations of multiple satellites with microwave instruments can help account for the reduced temporal resolution.

Slide 4: Typically, in the microwave part of the spectrum, we refer to bands by their frequency rather than wavelength. The frequency, denoted here with Greek letter nu, is just the speed of light, or about 3 x 10^8 m/s divided by the wavelength. Henceforth in this course, we will use frequency when referring to passive microwave and wavelength when referring to passive IR or visible remote sensing. Frequencies in the microwave are small enough so that we can re-express the Planck blackbody equation as a linear function of temperature. For microwaves (long wavelength), the ratio in the exponent in the denominator becomes very small. We can use the fact that the natural exponential function for small exponents simplifies to 1 plus the exponent to remove the exponent from the denominator. After doing some algebra to cancel like terms in the fraction, we are left with the following expression that states the microwave blackbody radiance is linearly proportional to temperature. Thus, the radiance *L* is just emissivity times the Planck radiance, and we can define a microwave brightness temperature that is simply emissivity multiplied by actual temperature. As we will soon see, this means we can substitute radiances for temperatures in Schwarzchild's equation. Note before we move on that in the IR, we often assumed that emitters on Earth, and especially the ocean, could be approximated as blackbodies. This is most certainly not the case in the microwave, and furthermore, the emissivity of the ocean changes as the properties of the ocean surface change. Thus, this emissivity term becomes very important for microwave remote sensing.

Slide 5: In the top right panel, frequencies associated with corresponding wavelengths are shown on the top and bottom x-axes. Between roughly 10 and 50 GHz, the atmosphere is relatively optically thin. Let's suppose, for an idealized case that we have a homogenous single-layer atmosphere with some temperature. The surface radiance is that from emission *and* reflectance, and sources along the path include emissions only.

Slide 6: The general form of Schwarzchild's equation is shown at top, with the top of atmosphere radiance on the left-hand side. If we suppose the atmosphere has depth H and a volume extinction coefficient of sigma_e, which is just the volume absorption coefficient in this case, then the optical depth is H*sigma_e, which we can use to define the direct transmissivity. We know that the portion of the top of atmosphere radiance emitted by the atmosphere is just the Planck radiance of the atmosphere multiplied by the absorptivity of the atmosphere, which through Kirchhoff's Law, is the emissivity of the atmosphere. This B times 1 minus tau also represents the downward emission of microwave radiation by the atmosphere that may later be reflected off the surface and back to space. We can then express the top of atmosphere radiance as a function of Planck radiance as follows. The first term represents radiation emitted by the surface that is transmitted all the way to the top of the atmosphere. Note the presence of the emissivity term, which is shown as a function of temperature and salinity. The second term captures the radiance emitted by the atmosphere—the B times 1 minus tau—that is reflected off the surface, which is the 1 minus epsilon, that is transmitted all the way through the atmosphere, which is controlled by the last tau term. The reflectance is 1 minus emissivity because the emissivity equals the absorptivity and all radiation not absorbed is scattered off the surface. The final term describes emissions from the atmosphere directly to space.

Slide 7: Using the linearization that we worked through earlier, however, we can convert the radiances to temperatures to express brightness temperatures into contributions from surface emissions, surface reflections, and atmospheric emissions. Two gases, water vapor and molecular oxygen, contribute to absorption of microwaves. Therefore, the direct transmittance of the atmosphere is dependent heavily on wavelength and water vapor concentration in the atmosphere. Like in the IR, we can take advantage of these absorption bands to garner information about profiles of temperature and humidity in the atmosphere using multispectral sounders.

Slide 8: While scattering of microwaves is not as prevalent as scattering of visible light, complicated scattering patterns can still occur in clouds, especially at high frequencies. We especially take advantage of this scattering in active microwave remote sensing, or radar, which we will discuss in the next lecture series. The equation currently shown at the top is an expression of the volume scattering coefficient. It is dependent on the size of scattering objects, the number of objects, and the scattering efficiency, represented here by *Q*. The scattering efficiency is depicted by the plot at bottom left, which shows how *Q* varies as a function of size parameter. Recall that the size parameter is related to the ratio of the circumference of a scatterer to the wavelength of radiation interacting with the scatterer. The scattering efficiency is a complicated function that also depends on the imaginary component of the index of refraction, which describes the absorption properties of the scatterer. We presume that a significant scattering interaction happens if the size parameter exceeds 1. For air molecules, at least this size parameter can occur for radiation at wavelengths of about 600 nm or less. However, for typical rain drops, which are much larger—around 1 mm in radius—radiation shorter than 6 mm can encounter significant scattering. This corresponds particularly with bands at frequencies greater than 50 GHz; however, larger rain drops cause appreciable scattering at even lower frequencies. The plot at right should be familiar from earlier in the quarter, which helps to visualize the differences between scattering of microwave vs IR or visible radiation.

Slide 9: We can then explore in more detail how clouds might impact microwave radiative transfer. We can effectively see through aerosols and air molecules at any microwave frequency, but our ability to see through clouds depends on frequency. The first column in the table shows frequencies on the Defense Meteorological Satellite Program's Special Sensor Microwave Imager, or SSMI. The corresponding wavelengths for each frequency are shown in the second column. The critical radius of a scattering object is given in the right-most column. It is the wavelength of radiation divided by 2-pi, the radius required to obtain a size parameter of 1. The 91 GHz band encounters significant scattering when it interacts with objects that are at least half a millimeter in radius. Plenty of liquid water drops in precipitating clouds are near or exceed this size. As we move to lower frequencies, larger objects are required to achieve a significant scattering interaction. The 19 GHz band requires an object that is half a centimeter in diameter to cause a significant scattering interaction. This would be a large raindrop. Therefore, low frequencies that are not heavily absorbed are generally able to see some properties of the surface. However, because the atmosphere is not completely transparent at any of the frequencies listed, any brightness temperatures would contain contributions from both the surface and atmosphere. The higher frequencies, however, are more likely to not see emission from the surface because that radiation is exposed to both scattering and absorption in the atmosphere.

Slide 10: The upshot of this is that as hydrometeors in a cloud increase in size, the optical depth along a path through the cloud to a satellite also increases. We can insert various hypothetical values of atmospheric temperature and transmissivity into our equation for brightness temperature to see the impact of clouds on microwave brightness temperature over ocean. Consider the hypothetical single-layer atmosphere shown at left. The surface temperature is 280K, and the air temperature is 260K. We'll assume the cloud on the left is optically thin and the direct transmittance is 0.9. The cloud on the right is optically thick so that the direct transmittance is zero. Very importantly, suppose the emissivity of the ocean is 0.4, a typical value, which is also dependent upon frequency. If we plug these values into our equation, we see that the optically thin atmosphere has the lower brightness temperature. In fact, the brightness temperature is well lower than the temperature we expect any surface on Earth to have. Thus, *at low microwave frequencies, low brightness temperatures typically correspond with few or no clouds.* This is contrast with IR brightness temperatures, which generally are largest in a cloud-free environment. The figure on the right shows the implication of this on estimations of rain rate from microwave brightness temperatures. Low brightness temperatures generally correspond with near-zero rain rate over ocean. However, because land is much more emissive of microwaves than water, we expect to see high brightness temperature over land in cloud-free areas. Over the ocean, as we continue to the right on the x-axis, rain rates will eventually increase as brightness temperatures decrease. This is because eventually, no radiation from the poorly emitting ocean reaches the top of the atmosphere, and only the emissions from near the tops of clouds reach space. Frequencies that are more susceptible to scattering and absorption, like 85 or 91 GHz radiation generally have brightness temperatures that decrease as cloudiness increases for all but very dry atmospheres. However, lower frequencies can be less straightforward to interpret. As you can see, a 240K brightness temperature at 18 GHz over ocean, for example, could correspond to a rain rate of either 10 mm/hour or 50 mm/hour. We will address this ambiguity that arises in interpreting brightness temperatures over ocean in a future module.

Slide 11: Finally, this rather complicated plot shows the effects of various surface or atmospheric properties on brightness temperature at 3 SSMIS channels, 19, 22, and 37 GHz. The x-axis is frequency. The y-axis represents the change in brightness temperature that occurs as a result of a change in one of the properties plotted. The plus and minus signs represent the sign of this change in brightness temperature. For example, the yellow line represents the effect of water vapor on brightness temperature. It is always above zero on the y-axis, meaning that increases in water vapor cause increases in microwave brightness temperatures at all frequencies shown on the x-axis. There is a peak in sensitivity of brightness temperature to water vapor at 22 GHz, which is located in a water vapor absorption band. Therefore, a 22 GHz sensor will detect less radiation emitted by the surface because much of the surface radiation is absorbed and re-emitted by water vapor. Since the satellite sees more radiation from the atmosphere, which is more emissive than the ocean, the brightness temperature is higher as water vapor concentration increases. Likewise, brightness temperature generally increases in all bands as liquid water concentration increases. We can tell that 37 GHz radiation is more sensitive to liquid water than 19 GHz radiation, consistent with our discussion of scattering from before. Brightness temperature is sensitive to salinity primarily at frequencies less than 10 GHz because ocean emissivity at those frequencies is dependent upon salinity. Brightness temperature generally increases as wind speed increases at all frequencies because the rougher ocean surface is a little more emissive. Finally, the temperature of the ocean surface effects its emissivity. The effect of surface ocean temperature on brightness temperature varies in a complicated way, often changing sign as a function of wavelength. For the frequencies shown, emissivity generally decreases as frequency increases. In fact, surface temperature will increase the 19 GHz brightness temperature but decrease the 22 GHz brightness temperature. If the differences in brightness temperature between bands due to liquid water or water vapor (and perhaps wind speed) can be constrained, various bands can be used together to help estimate surface temperature. Passive microwave observations are indeed used in concert with infrared radiances and surface data to create global analyses of sea surface temperature as well as other variables as we will see in subsequent modules.

Module 4.2

Slide 1: Next, we will become a little more familiar with the Special Satellite Microwave Imager/Sounder, or SSMIS. In particular, we'll focus on the Imager and take a first look at polarization of radiation.

Slide 2: Radiation appears in two basic polarizations: horizontal and vertical. The polarization refers to the propagation of radiation through an electromagnetic field. Vertically polarized radiation (in some frame of reference) has its electric field oriented vertically, like shown in the image at the top. The magnetic field runs perpendicular to the electric field. As seen in the animation on the right, when radiation in one horizontal dimension is emitted out of phase with its propagation in the other horizontal dimension, then the radiation is circularly polarized. Radiation can take on left-hand or right-hand circular polarizations as seen in the animation, which demonstrates the propagation of four different types of polarization. The circular polarizations can be defined as either clockwise or counterclockwise depending on whether defined from the point of view of the emitter or the observer.

Slide 3: As we mentioned in the previous module, the emissivity of the ocean changes as a function of many factors, including its temperature, salinity, the frequency of radiation emitted, or the roughness of the sea surface. It is also dependent upon the polarization of the radiation emitted. Objects including the land or ocean surface emit radiation simultaneously with various polarizations. In this figure we see how emissivity of the ocean varies simultaneously with angle of incidence from a smooth, flat ocean surface; frequency, with two different frequencies shown; and polarization, with the top two lines denoted vertical polarization at the two frequencies. An incidence angle of 0 degrees means that radiation is transmitted vertically. We see that in this case, emissivity is not dependent upon polarization, but that the emittance at 60 GHz is about 30% more than it is at 10 GHz. As the angle of incidence increases, the vertically polarized radiation is more effectively emitted, while the emittance of horizontally polarized radiation gradually decreases. Typically, microwave remote sensing is done at incidence angles of less than about 60 degrees.

Slide 4: Next, we'll work through an old example of microwave brightness temperature observed in a wintertime scene from the late 1900s. These were observed by SSMI before the sounder was added. The imager operated in four frequencies—19, 22, 37, and 85 GHz—and in orthogonal polarizations. These are similar to the imager on SSMIS, which operates today and is the focus of the lab on passive microwave imagery. It operates at the same frequencies, but the 85 GHz band is replaced by a band at about 91.6 GHz—which is sometimes referred to as both the 91 or 92 GHz band. SSMIS also detects emissions in the oxygen band near 55 GHz. The image shown at top left is an example of 19 GHz vertically polarized brightness temperatures observed over the northeastern portion of the Soviet Union and nearby ocean. In short, we can call this the 19V channel. Brightness temperatures over the ocean are generally low, while brightness temperatures over land are relatively warm. The yellow line near the top of the image is the northern shoreline of the landmass, and the warm temperatures to the north of the coast are emissions from sea ice.

Slide 5: In comparison—and note change in the color scale—the 19H channel shows lower temperatures over the ocean than in the vertically polarized channel of the same frequency. Two locations over the ocean, which were also faintly seen in the 19V channel, show slightly elevated brightness temperatures. These may indicate locations where the wind is roughening the surface or locations where cloud is “replacing” some of the emission from the poorly emitting surface with emissions for the relatively good emitting cloud.

Slide 6: As we move to 22GHz, the contrast between land and ocean remains, but some clouds become apparent. Some of the warmer brightness temperatures over ocean and lower temperatures over land might be caused by cloud. Emissivity of 22 GHz radiation is the same in the vertical and horizontal polarizations, so only a vertical channel is used.

Slide 7: At 37 GHz, sea ice, which is a relatively good emitter compared to liquid water, has the highest brightness temperature in this scene. The ocean brightness temperature is higher than in 19 GHz because the ocean is more emissive at 37 GHz, but it is lower than at 22 GHz, because more surface emissions are absorbed by oxygen in the atmosphere. At 37 GHz, clouds over the ocean begin to become more apparent, and some spots with lower brightness temperatures become evident over land depending on the emissive properties of the land surface in that particular location.

Slide 8: At 37H, noting again that the color scale changes, we see again that brightness temperatures are generally cooler than in the vertically polarized channel.

Slide 9: Finally, when we look at higher frequency data—in this case at 85 GHz—we are first able to notice that the spatial resolution is much higher than before. Individual cloud features aligned in a band are clearly visible offshore. The clouds are shallow. We know this because their brightness temperatures at 85 GHz are warmer than the ocean, which is still a poor emitter. However, because high frequencies are particularly susceptible to absorption by water vapor, the 85 GHz brightness temperature in cloud-free areas over the ocean is higher than it was in lower frequencies. Again, this is because fewer emissions from the poorly emissive ocean are able to reach space unaffected by atmospheric constituents. We also notice that the land has a much lower brightness temperature at 85 GHz than at lower frequencies. Here, the brightness temperature is no more than 200 K, while in the 19V channel, it was 240–250K. Emissivity of the land generally decreases from 19 to 85 GHz; however, the land in this case may be covered partially by snow, which has emissivity that can vary as a function of the snow density as well. Most fundamentally though, a lot of the variability you have seen in brightness temperatures is related to variability in emissivity of the surface, or variability in the amount of radiation emitted by the surface that is able to pass completely through the atmosphere.

Slide 10: And finally, we can look at the 85 GHz H channel, which highlights many of the features shown on the previous slide, although a different color bar is used to highlight some of the variability over land, much like we saw at 37 GHz.

Slide 11: We'll spend a little bit of time now looking through some of the specifics of data collection as pertaining to SSMIS. It, like many other low-Earth orbiting satellites, is in sun-synchronous orbit. The equatorial crossing times vary by satellite and evolve over time but are generally in the afternoon and on the ascending node. The scan geometry of the sensor is shown here. It scans in a cross-track configuration at an angle of 45 degrees off nadir, meaning that it sweeps back and forth in arcs along the surface that are perpendicular to the flight track. A sweep back and forth is denoted by the curved black lines. As the satellite moves along at an altitude of over 800 km, it "paints" in a swath along its track; the width of that swath is about 1700 km. Within that swath, it collects data at various frequencies previously mentioned. Therefore, the microwave sensors, unlike the geostationary sensors, only observe a small location relative to geostationary instruments over the same time period.

Slide 12: Because the emissions of microwaves by Earth are low compared to IR, the spatial resolution of microwave data, despite the sensors being closer to Earth than, for example, GOES, is lower than geostationary data or low-Earth orbit visible and IR data. The spatial resolution depends on frequency, with the higher frequencies that are a little closer to Earth's peak Planck emittance having higher spatial resolution. The three ellipses here are shown to scale relative to each other. A larger footprint is required at low frequency channels such as 19 and 22 GHz; therefore, data is rather coarse. However, a relatively small footprint can be used at 92 GHz, which is similar to the 85 GHz channel on the old SSMI. You can see the ramifications of the different spatial resolutions by looking back at the previous slides of sample data.

Slide 13: SSMIS also uses a sounder to gather information about profiles of temperature and moisture. The fundamental mechanism for how these quantities are derived is the same as for other infrared sounders or microwave sounders, like those on JPSS platforms. The SSMIS sounder bands are located near two absorption bands. The first is the oxygen absorption band near 60 GHz, around which several bands using some different polarizations are located, and the other is a water vapor absorption band close to 200 GHz. The two sets of bands are useful for estimating, respectively, profiles of temperature and humidity in the atmosphere.

Slide 14: Approximate weighting functions of the modern-day SSMIS bands are shown here. On the left are the weighting functions for channels surrounding the oxygen absorption band. These allow for estimation of a crude vertical profile of temperature in both the troposphere and stratosphere. On the right are the average weighting functions of the imager channels, as well as the bands centered well above 100 GHz. These are more useful for deriving profiles of water vapor in the troposphere.

Slide 15: Solid vs dashed lines in this figure show differences in some of the upper atmospheric weighting functions for polar vs tropical regions.

Slide 16: Several products can be derived from the SSMIS brightness temperatures. One half-day's worth of swaths for one satellite are shown here. This figure shows column-integrated water vapor derived from the multiple bands. It represents the mass-weighted integral of water vapor present in a column. The red colors represent large values of water vapor, which are mostly present in the deep tropics.

Slide 17: Other products include surface wind speed, where for example, we can see the magnitude of winds in mid-latitude cyclones pop out.

Slide 18: Cloud liquid water

Slide 19: Rain rate, which is closely related to cloud liquid water in many locations.

Slide 20: Products like this map from the University of Wisconsin are derived for the entire globe hourly. These use data from a combination of different sensors, including those aboard the various DMSP satellites, to create more continuous estimates of atmospheric humidity. A network of various microwave sensors aboard numerous platforms substitutes for the lack of temporal resolution that a single satellite can provide at any given location. This instantaneous, multi-satellite estimate of atmospheric water vapor looks similar to…

Slide 21: the monthly average derived from a single satellite. It shows, again, how more moisture is detected in the tropics. We are able to estimate this by leveraging the extinction of radiation in the atmosphere that is emitted by the surface. For example, in two equally warm environments, and using a frequency that is sensitive to water vapor absorption—which is essentially all bands except those in the oxygen absorption band—the moister environment will often yield higher brightness temperatures over ocean, telling us that more water vapor is present.

Slide 22: This corresponds nicely with the monthly averaged estimated rain rate from the same single satellite and provides some insight into a key aspect of tropical dynamics—the relationship between tropospheric moisture and moist convection.

Module 4.3

Slide 1: When viewing microwave observations in real time, ambiguities can arise in how to interpret them. For example, one value of brightness temperature could correspond to two very different atmospheric states. In this module, we will address these ambiguities and discuss how to use multiple bands cooperatively to interpret microwave data.

Slide 2: We'll look at a couple of oversimplified atmospheric states to demonstrate the point then look at some real data. Let's first suppose that we have an idealized homogenous atmosphere that has the same temperature throughout its entire depth. We'll make that temperature 260K and make the surface temperature 300K. We will assume that we are over ocean and that the emissivity of the ocean varies by frequency as shown at the bottom, with emissivity increasing as frequency increases over the range of SSMIS channels. The equation for estimating brightness temperature for a homogenous atmosphere is shown at center, and we'll fill in the lines in the chart at the top of the table. We will consider a completely dry, cloud-free environment, a "moderately dry", cloud-free environment, a moist environment with some shallow clouds, and a moist environment with deep convection. These environments will have different optical depths, and the optical depths will also vary by frequency.

Slide 3: Suppose those optical depths are like those shown in this table for the four different environments and four different frequencies. We'll show an example of the moderately dry environment at 22 GHz first.

Slide 4: We can plug in values for emissivity at 22 GHz, the appropriate temperatures, and optical depth to get the brightness temperature.

Slide 5: If I've done the math correctly, we get about 232K.

Slide 6: We can follow the same exercise for all of the combinations of optical depth and emissivity—given the same atmospheric and surface temperatures—and come up with the values shown at the top. Generally, brightness temperatures are warmer at 92 GHz than at other channels because the 92 GHz channel is very sensitive to water vapor, and when cloud drops are present, is highly susceptible to scattering as well. This means that the emissions from the poorly emitting ocean do not reach space. Likewise, as the optical depth increases for each channel, the brightness temperature increases. Note, though, this is for an unrealistic single-layer atmosphere.

Slide 7: A more realistic atmosphere is more complicated. Our equation that assumes a homogenous atmosphere does not work anymore. Instead, we must integrate the last two terms over the optical depth of the atmosphere as a function of height.

Slide 8: I've selected a few possible values of the "representative" temperature of the atmosphere by assuming where the peak of the weighting functions for each band might be located in the various environments. We'll plug these air temperatures into our linear equation to create a first guess of what the brightness temperature might be in a more realistic setting. Note that in the completely dry environment in which direct transmittance is 1, the atmospheric temperature is unimportant.

Slide 9: When we do this, we get the brightness temperatures shown at top. The channels for which the atmosphere is more transparent, like 19 and 37 GHz, generally see an increase in brightness temperature relative to the homogenous atmosphere, while the 22 and 92 GHz bands experience a decrease in brightness temperature in the optically deep environments such as when deep convection is present.

Slide 10: We can sketch how brightness temperature evolves as a function of water vapor concentration or optical depth. For the homogenous atmosphere, the brightness temperature eventually asymptotes to the atmospheric temperature for all frequencies if enough vapor or liquid water is present to absorb or scatter all surface emissions. However, for the more realistic atmosphere, brightness temperatures first rise as vapor increases but then fall again. This happens more quickly as a function of water vapor concentration at frequencies more susceptible to water vapor absorption or scattering by hydrometeors, such as 92 GHz.

Slide 11: What this means to an observer is that one brightness temperature could represent two very different atmospheric states. In this case a 190K brightness temperature at 92 GHz could represent a totally dry atmosphere or deep convection.

Slide 12: This is reminiscent of the figure shown previously in this lecture series, showing how rainfall estimates over ocean changed as a function of frequency and brightness temperature. Lower frequencies experience some ambiguity as well, meaning that I didn't select the most realistic values of the "average" atmospheric temperature for emissions in our more realistic example. For example, a 240K brightness temperature at 18 GHz could represent a light or heavy raining state. However, if we consulted the 91 (or 85) GHz brightness temperature at the same time and location, it would be much higher for the lighting raining event than for the heavy rain. This helps us quickly overcome the ambiguity and interpret data in real-time.

Slide 13: An example of such ambiguity over land is shown here in GMI brightness temperatures at 10, 19, and 37 GHZ. Radar reflectivity for an afternoon in Texas is shown at top left. It shows a few cells containing some locally intense rainfall. In the GMI data, the cells are apparent within the white circles. However, especially at low frequencies that can see more surface emissions even in heavy rainfall, the difference between lakes—which are poor emitters—and the storm cell is small. By just looking at the 10 GHz brightness temperature, identifying lakes from storms could be challenging. Note that the emissivity of the land is generally much higher—close to 0.9—and so the brightness temperature over clear-air land is higher. Incorporating the 37 GHz data, however, yields some better insight into which of the "cool" areas in the 10 GHz image are actually convection.

Slide 14: A similar example from SSMIS is shown over the ocean in the Arabian Sea. Shown currently is 91 GHz brightness temperature. The deep convection is obvious and is denoted by the red colors. However, more optically thin convection that surrounds the red is denoted by green and yellow. In the far left of the image, green and yellow colors at the mouth of the Persian Gulf look the same as the optically thin cloud.

Slide 15: The 37 GHz brightness temperature also identifies the cloudy regions as areas of relatively warm brightness temperature, but the mouth of the Persian Gulf now has a very low brightness temperature. This tells us that the mouth of the Persian Gulf is actually cloud-free. It is probably a particularly dry area over which even the 91 GHz channel was able to detect some surface emissions.

Slide 16: Another product available operationally is the PCT product, which stands for polarized corrected temperature. The general formula for PCT is shown at bottom and is a weight of the horizontally and vertically polarized brightness temperatures in the same frequency. The capital theta is a coefficient that must be determined prior to the product being generated; as a forecaster, you would not be concerned about what the coefficient is. The PCT product will generally highlight just the deep convection. Note that the color scheme varies depending on the source of your data.

Slide 17: Finally, we can return to our example over Texas, and compare vertically polarized brightness temperatures in several bands—now including 89 GHz—to the PCT product for each channel. The lakes that were especially apparent in the low frequency imagery have now disappeared, and the convection is highlighted. Still, because the atmosphere is so transparent the low frequency microwave, consultation of multispectral data—instead of just a single channel—is prudent to accurately interpret the scene.

Module 4.4

Slide 1: One of the many applications of passive microwave is determination of surface wind speed. This module will briefly look into how wind speed is derived from microwave brightness temperatures, and we will also take a look at a passive radiometer that can estimate surface wind direction in certain conditions as well. The background image is from a previous module and shows an example of surface wind speeds derived from a passive sensor.

Slide 2: As we've seen before, the emissivity of the ocean surface depends on the incidence angle of the emitted radiation and the frequency at which the radiation is emitted. The emissivity is also dependent upon the polarization, with emissivity of vertically polarized radiation increasing as incidence angle increases, and that of horizontally polarized radiation decreasing. Wind speed dependence also exists and is a function of incidence angle as well as seen in the panels to the right. The pink box represents incidence angle at which typical passive microwave radiometers operate, and the figures show typical brightness temperature at 19.4 GHz as a function of incidence angle and wind speed. The dashed line represents a smooth, calm ocean, and solid lines represent various wind speeds. At typical viewing angles used by passive microwave instruments, emission of vertically polarization radiation by the ocean does not change as a function of wind speed. However, horizontally polarized radiation is emitted more as wind speed increases. If other properties of the sea surface and atmosphere are known, this fact may be leveraged to derive wind speed.

Slide 3: Such sensors can estimate wind speed based on brightness temperatures in vertically and horizontally polarized channels; however, they are not able to retrieve wind direction unless fully polarimetric and under certain conditions—such as high wind speeds and with low cloud cover. Only frequencies that can reliably see the surface are used in passive wind speed retrievals. A high frequency channel such as 92 GHz would detect few emissions from the surface in places where water vapor concentration was high, such as the tropics; therefore, it would be useless for deducing surface properties.

Slide 4: WindSat was designed by the Naval Research Laboratory Remote Sensing Division and launched in 2003 as part of the Coriolis satellite. It is still active as of 2020 and is the oldest multispectral polarimetric microwave radiometer in orbit. WindSat operates at five frequencies, which we will see momentarily. Shown here is the ocean emissivity as a function of wind speed for one of those channels: 6.8 GHz. The top and bottom sets of lines represent horizontally and vertically polarized emissions, respectively. The y-axis denotes the change in emissivity caused by wind relative to some background state with no wind and a fixed sea surface temperature. A strong 30 m/s wind would increase the ocean emissivity for 6.8H by about 8% and 6.8V by about 5%. Generally, emissivity increases as a function of wind speed for both polarizations; however, at low wind speeds, ocean emissivity of vertically polarized radiation experiences a small decrease before again increasing.

Slide 5: The five frequencies used by WindSat are 6.8, 10.7, 18.7, 23.8, and 37.0 GHz. Recall from a previous module that the 6.8 GHz channel is quite sensitive to SST and the 23.8 GHz channel sits along a water vapor absorption band. These two frequencies can therefore be especially used to determine the effect of SST and tropospheric humidity on brightness temperature, while the others are then used for estimating the wind speed. Because the lower frequencies are not as heavily scattered by rain, WindSat can provide some estimates of coarsely resolved wind speed even in places where precipitation occurs. Full polarimetry, which involves detecting vertically and horizontally polarized radiation in addition to slant polarization and left- and right-hand circular polarization, is used on WindSat. Passive observations can be used to determine wind speed in one of two ways: 1) By detecting horizontal and vertical polarization from various angles, or 2) obtaining a fully polarimetric single view of the scene, which is limited at estimating wind direction where surface wind is light.

Slide 6: An example of WindSat wind speeds during an incomplete day of observations is shown here. WindSat is in sun-synchronous orbit so ask yourself, is this a set of descending or ascending orbital data shown?

Slide 7: By zooming into some sector, we can now see wind vectors derived from the fully polarimetric data.

Slide 8: An additional, but very different, method of detecting surface wind speeds over ocean is using bistatic scatterometry. This involves the use of a satellite instrument that, unlike traditional scatterometers, does not transmit its own signal. Instead, it detects reflected GPS transmissions off of the surface. Given that the GPS satellite location is known, and if we know the power transmitted by GPS, we can use scatterometry, the focus of a different module, to estimate both wind speed and direction without need to transmit a different signal. One limitation of this method is that if the GPS transmitted power changes unexpectedly, the retrieval to estimate wind speed must be recalibrated. The Cyclone Global Navigation Satellite System, or CYGNSS, is a current example of bistatic scatterometry. The GPS signal, which is in L-band, is not impacted by water vapor or liquid water. Therefore, unlike traditional scatterometers—or even passive radiometers in some cases—CYGNSS satellites can detect wind speed even in heavy rain, such as in tropical cyclones from low-Earth orbit.

Slide 9: Just to visual this, a CYGNSS satellite consists of only a receiver that detects some radiation that is scattered off the surface by some GPS transmitter in semi-synchronous orbit. Not shown is the full CYGNSS constellation; it consists of 8 identical satellites.

Slide 10: One example of CYGNSS data is shown here in an Atlantic tropical cyclone for the left panels. For each panel, the blue line represents aircraft-based microwave estimates of surface wind in a cross-section through the storm, and the green line represents what the CYGNSS constellation of satellites would see in the same location. Generally, the two agree closely. The panel on the right shows several passes through a West Pacific typhoon over a 3-hour period, with warmer colors indicating higher wind speeds. Strong winds can be seen at the center of the tropical cyclone, which is difficult for traditional scatterometers to do effectively because signal is at least partially attenuated by scattering by liquid water. Like active sensing scatterometers, the radiation detected is reflected off of a large area, and the wind speed retrievals are somewhat coarse in spatial resolution, typically about 25 km by 25 km.

Module 4.5

Slide 1: A special application of microwave used for estimating surface wind speeds and directions over ocean is scatterometry. It is an active sensing technique, meaning that it involves transmission of a signal from space that is reflected off the ocean surface and returned to the satellite sensor. This module discusses the basic mechanisms through which scatterometry works.

Slide 2: So far in this course, we have learned about passive sensors, those that detect radiation that is scattered off of or emitted by Earth's surface or atmosphere. Active sensors transmit a signal and used the returned power from that signal to infer some property of the surface.

Slide 3: Scatterometers measure backscatter off the ocean surface. Emissions are no longer very important because the transmitted signal is stronger than the small emissions of the ocean surface at microwave frequencies. As we will see, the backscatter changes as a function of the surface roughness of the ocean, which depends on—and generally increases with—wind speed.

Slide 4: First, we want to define a quantity called the scattering cross-section. At bottom right is a diagram containing a scatterometer at top. As we will discuss soon, the instrument would likely not scan straight down, but this is just for illustrative purposes of the scattering cross-section. The signal transmitted is represented by the concave up, black lines, and the returned signal is denoted by the concave down, red lines. P_T is the transmitted power, and P_R is the received backscattered power at the sensor. The top-most equation is the power flux density of returned power at the antenna. G is the antenna gain, which we will discuss in more detail in the next lecture series on radar, and r is the range of the reflected signal—in this case the altitude of the satellite. The total received power is a function of the effective scattering area at the surface, or A_{eff}. The scattering cross-section is the integral of the normalized scattering cross-section, or σ_0, over the area subtended by the transmitted signal. We can use this normalized scattering cross-section to derive wind speed and direction because it depends on the properties of the ocean surface. In other words, everything in our equation for received power at the antenna is fixed except of the backscatter cross-section, so by converting power received to some value for σ, we can back out information about the wind because it predominantly impacts the scattering properties of the ocean surface.

Slide 5: There are two primary types of scattering that impact σ_0 very differently. The first is called specular reflection. It is mirror-like and occurs when the sea state is very calm. It dominates when the incidence angle—or the angle off of zenith—of the transmitted radiation on the sea surface is less than about 15–20°. In the example on the right, the transmitted signal interacting with the ocean is shown in four locations. In three, the surface is nearly perpendicular with the signal, and most of the transmitted radiation is scattered back to the sensor. In contrast, the red lines—third from left—indicate a location where specular reflection does not occur because the path of the beam is not orthogonal to the ocean surface locally. The plot on the left shows us how σ_0 varies a function of incidence angle and wind speed. The orange line is some unspecified wind speed, and the blue line indicates a new σ_0 after increasing the wind speed. In the specular reflection regime, the backscatter cross-section tends to decrease as the wind increases for small angles.

Slide 6: The other type of scatter is resonant scatter, which is also called Bragg scatter. It occurs when some component of ocean waves has a wavelength that is equal to half of the scatterometer wavelength divided by the sine of the incidence angle. The *n* in the numerator is any positive integer that denotes that a family of Bragg-scattering solutions exist. Bragg scatter occurs when the backscatter of two successive waves of the wavelength λ_w are in phase with each other, thus increasing the power returned to the sensor. For small values of n, this generally involves short-wavelength wind-driven capillary waves since the wavelength of the transmitted radiation is on the order of a few centimeters.

Slide 7: Looking first at the panel on the left, at angles greater than about 25–30° and less than about 65°, Bragg scattering causes an increase in the backscatter cross-section as wind speed increases. The dashed black line next to the orange line is an example of backscatter cross-section as a function of incidence angle. The orange line denotes the new backscatter cross-section given some hypothetical unspecified increase in wind speed. Therefore, at this range of incidence angles, we are able to unambiguously determine, given some transmitted signal, that the wind speed increases as the power returned to the sensor increases. For this reason, most scatterometers scan off nadir in the range where Bragg scattering is prevalent. Shadowing and refraction of radiation prevent scatterometers from operating at incidence angles much larger than 65°. To further demonstrate the point, the panel on the right shows for various incidence angles how the backscatter cross-section typically changes as a function of wind speed. For low incidence angles, an increase in wind speed decreases the cross-section, while for incidence angles of 20° or greater, the opposite is true, and wind speed causes increasing backscatter.

Slide 8: Shown are some potential scattering patterns in orange for a given transmitted signal in blue for a variety of sea surface states. The incoming signal is at a steep angle so that it can detect backscattered Bragg reflection. For the smooth surface in the top left, specular reflection dominates, and most of the radiation is forward scattered away from the sensor. In this case, a scatterometer would receive little to no power from the transmitted signal and could deduce light winds. XXXXXX As we move from top right to the bottom row, the wind speed increases, and the roughness of the surface increases as well. With more Bragg scatter occurring, the returned power to the sensor increases, and the increased power can be interpreted as increased wind speed.

Slide 9: Scatterometers also report wind direction. To do so requires viewing the same scene from two or more angles. Suppose we have some sort of scene like shown on the left, with the wind blowing from bottom to top. The wind-driven wave crests and troughs will tend to be oriented along the direction of the wind flow. If a scatterometer is upwind or downwind of the scene, it will detect maximum backscatter via the mechanisms described before. The returned power will be slightly larger if viewed from downwind because capillary waves and foam formation are preferred on the downwind side of wind-driven waves. If the scatterometer views the scene at a direction perpendicular to the wind, the backscatter cross-section is relatively small, and so is the received power. When the scatterometer detects a single value of returned power, it does not know what the wind direction is. We are only able to derive some value of σ_0. Let's suppose an instrument receives power consistent with a backscatter cross-section of -11 dB, as denoted by the blue dashed line on the right figure. We don't know the azimuthal angle that the transmitted signal is relative to the wave crests and troughs. We could be observing a strong wind speed that would increase σ_0 at an angle orthogonal to the waves that would decrease σ_0 or we might be looking at a lower wind speed at an angle closer to parallel to the waves. Each black line on this panel represents a family of possible backscatter cross-sections and azimuth angles relative to upwind. By following the blue line across at -11 dB, we see it crosses two black lines, plus an infinite number of lines not shown between them. We could, for example, be looking at 7 m/s wind from upwind, a 7.5 m/s from downwind, or a 10 m/s wind from cross-wind.

Slide 10: Given just a single view from some angle θ, we could come up with a solution like this, now shown as a plot of wind speeds as a function of wind direction given some value of σ_0. By itself, this is rather useless because any wind direction is possible, and wind speeds of between 16 and 28 m/s are possible. Another view angle is required.

Slide 11: If we add a second view, then the only possible solutions are where the two curves intersect. They do so in four locations, greatly reducing the ambiguity.

Slide 12: By adding a third, or even fourth, view, we further reduce the ambiguity and close in on a solution. Many scatterometers only use 2 or 3 view angles and rely on observations and model analysis to discard one or two of the possible solutions.

Slide 13: The Advanced Scatterometer, or ASCAT, is aboard the EUMETSAT MetOp series of satellites. It scans in two off-nadir swaths at the same time, scanning at an angle of 29.3° off-nadir to the west and east, while not scanning the sub-satellite point at all. In each swath, three signals are transmitted: One at 45° ahead, one perpendicular to the flight track, and one at 45° behind. As the satellite moves along its track, all three beams observe the scenes in each viewing swath. This provides views from three azimuthal angles, which helps reduce or eliminate ambiguity.

Slide 14: Some more information on various scatterometer platforms—past, present, and future—can be found at the bottom link. Some real-time data can be accessed at the top link.

Slide 15: An example of data taken from the first link on the previous slide is shown here. This is from the ASCAT instrument on MetOp-A. Warmer colors indicate stronger winds, but we will have to zoom in to see wind direction. A single scan along the ascending node is circled in black. Note that at the center of the ellipse, an empty sector is present, which occurs at near-nadir angles where Bragg scatter would not contribute to the backscatter.

Slide 16: We can zoom into a region to see wind vectors, which in this example, are shown for off the California coast. In this example, we see northerly to northeasterly surface winds, with strongest wind speeds found at this time just offshore northern California at the top of the image. Again, areas with missing data are either within 29.3° of nadir or between successive passes, each farther to the west than the previous. Although not shown here, scatterometer data can be “rain-contaminated”. This happens when liquid water scatters a large amount of the transmitted signal, weakening the return signal to the sensor. Such vectors are usually denoted somehow on real-time data, although they are more common at higher frequency scatterometers—such as those that operate at K_u-band near 13.4 GHz—than those like ASCAT, which operate in C-band near 5.25 GHz and are less susceptible to scattering.

Module 4.6

Slide 1: This module provides a brief overview of altimetry, an active remote sensing technique used to estimate sea surface height, including significant wave height. Over long periods of time, altimetry data can be used to observe changes in sea surface height caused by anthropogenically forced global warming.

Slide 2: Altimeters transmit at nadir toward the surface, and we are interested in knowing the time delay between transmission and receipt of the backscattered return if we want to compute sea surface height. The pulse width is very short, and the remainder of the time between transmission and return is spent listening for the return signal. This takes roughly 4 to 5 milliseconds; then a new transmission occurs. Given a known speed of light, the time delay tells us the distance between the satellite and the surface. Variations in the ocean topography can be caused by many factors: Tides or atmospheric conditions can alter the surface. More importantly though, the uneven distribution of mass in the planet impacts the gravitational pull globally, which also alters the mean sea surface height. The topography of the ocean surface given Earth's gravitational field is known as the geoid. This is essentially the background ocean topography in the absence of any other forcing. In order to estimate the sea surface height to the intended accuracy of 2 to 3 centimeters, the geoid must be known as accurately as possible, and the height of the satellite must be known exactly. The latter can be constrained using GPS, while the former has been mapped during several prior missions. The H on the Earth shown represents a region with a higher geoid. Therefore, we would expect the typical time delay here to be smaller and would need to account for this when determining sea surface height above the background state.

Slide 3: The geoid relative to a reference ellipsoid shape is shown here. The undulation of the geoid is the variation between it and the reference ellipsoid, typically about 60 meters or less. The sea surface height, denoted by the dashed line, is the variable of interest, and the time delay to it is what is measured by altimeters. The difference between the sea surface height and the geoid is the topography.

Slide 4: Orbital characteristics for one altimeter, Jason-3, are shown here. Note that the orbital altitude is higher than most other low-Earth orbiting satellites we have discussed in this class. This is because at lower altitudes, even slight drag will cause the orbital height to deviate more than is acceptable to achieve 2–3 cm accuracy.

Slide 5: Other information can also be garnered from altimetry data. Ocean floor topography can translate to detectable sea surface topography, and ocean surface currents can be deduced as well. Another variable provided by some altimeters is significant wave height. It is the average height of the largest one-third of waves. Altimeters can deduce wave height by considering the length of time that a return pulse takes to fully reach the receiver. For example, take the drawing at bottom left for a hypothetical wave with some height. For this single wave, the altimeter will receive its first signal in the time that it takes for the signal to travel to and back from the top of the wave. The entire backscatter from the wave is not received at once though. Signal that reflects off the bottom of the wave will take a little longer to reach the receiver in space. Let's suppose that time is Δt. If we double Δt, then we are doubling the likely wave height.

Slide 6: A daily composite of significant wave height derived from several altimeters is shown here. Notice how much of the surface is unobserved even by several sensors. This is because altimeters only scan in a narrow region directly beneath the satellite.

Slide 7: This is only one example of a data product publicly available. If we zoom in on Okinawa at this time, we see significant wave height along the swath passing over the island. Heights range from near zero to near ten feet along the swath seen in this image. They are represented by the blue and black numbers that could be different colors if the wave height was larger. The colors of the numbers alongside the image in this example—such as 0147—indicate the satellite that made the observation, and the number represents the time of the observation.