Course Name: An Introduction to Climate Dynamics, Variability and Monitoring

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APPLICATION OF VNIR IMAGING IN CLIMATOLOGY AND THERMAL INFRARED IMAGING SYSTEMS

Good morning class and welcome to our lectures in climate dynamics, climate variability and climate monitoring. So, we will continue our discussion on some of the applications of VNIR imaging, visible and near infrared imaging techniques. We looked at land vegetation and cover mapping, ocean color mapping. The other one that we will look into is cloud cover and aerosol concentration mapping. So, VNIR imaging is regularly used to delineate and track cloud cover, its extent, its height and its type. By tracking the motion of clouds, the wind speeds can also be estimated. Furthermore, aerosol concentration and type can also be estimated over dark surfaces like over oceans by identifying the contribution to the measured radiance from radiation scattered by the suspended material. So, the aerosol structure scatters radiation over oceans, for example, and that scattering can be detected, this scattering of the shortwave radiation can be detected by the VNIR sensor and helps us to identify the aerosol type, the aerosol concentration, the aerosol distribution. This is easier to do over ocean surfaces which have a very uniform emissivity. So we know what a clear ocean VNIR spectra will look like.

A reflection from an ocean VNIR spectra is clear, on a clear day will look like is well understood. So the aerosol structure can be inferred from the deviations of the, from the standard VNIR spectrum over a clear ocean. Okay. More recently more sophisticated algorithms have also been made it possible to find aerosol concentration values over through VNIR imaging over land surfaces as well, which is harder because land emissivity is highly variable and hence much more difficult to get standardized datasets for it. However, there are sophisticated algorithms and correlations that are used nowadays in modern satellite systems to find that and hence get aerosol concentrations for such cases. Clouds, especially cloud tops, reflect a lot of the VNIR radiation in the visible range. And hence, that can be used to find cloud, cloud distribution, track clouds, the height of the cloud, etc. and by tracking the cloud motion through successive VNIR images, we can also evaluate the wind speed at the cloud altitudes. For these cases, we need high resolution VNIR images. So, the 10 to 100 nanometer bandwidths are not sufficient, especially for aerosol concentration and distribution measures. Finely resolved VNIR spectra is used because aerosols will scatter only at certain spectra and identifying those spectra help us to detect aerosol concentration and aerosol types. So this data can again be seen from the PACE system because it also has an aerosol detector. And what we are looking here is the movement of aerosols, wind blown sand and dust particles moving from the Sahara Desert across towards the European continent through the Mediterranean Ocean. So, this side is Europe, this side is Northern Africa and this is the Sahara Desert. This is the true color image and you can see this optically also this aerosol is clearly seen.

Here we are looking at the aerosol optical depth. So, remember optical depth is directly correlated with the concentration of the aerosol along a column of the atmosphere. So, this optical depth at 0.55 micrometer wavelength. So, this is VNIR image at 0.55 micrometer wavelength, very finely resolved image and the optical depth can go up to 1. So, this region is entirely opaque by the large concentration of aerosols and it goes to 0 in other parts where aerosol is not present. So, based on this, the total density of aerosols between thus in the atmosphere can be obtained the concentration profiles. This is used to detect other things like the aerosol index as well as the albedo of these aerosols. So, single scattering albedo at 0.55 micrometers. So, this is the reflectivity or albedo of aerosol how much sunlight is reflecting back at 0.55 micrometers. And this again goes up to 1 in certain regions of heavy aerosol concentrations and goes to 0.8 at other regions.

So, this is the albedo of these aerosols at this wavelength range. And based on this, an effective aerosol layer height is also seen. So, how high up is the aerosol in the atmosphere? So, the layer height above the ground can also be obtained from this information. So, the top three panels of this OCI image depicting dust from North Africa carried into Mediterranean Sea show data that scientists have been able to collect in the past using satellite images. True color images, aerosol optical depth and UV aerosol index. So this is from the UV spectrum. So even the ultraviolet spectrum can be used to detect the aerosol concentrations. The bottom two images are novel to this PACE satellite system, and it is helping to create better climate models because it is giving the albedo of the aerosols. See higher the albedo in a certain aerosol region, less is the amount of sunlight going into the surface. So, that changes the energy balance of the world in that region.

So this helps to do climate modeling as well. Fraction of light scattered or observed which will be used to improve chemical models. Aerosol layer height tells us how low to the ground or high in the atmosphere the aerosols are, which aids in understanding the air quality as well. So, this kind of gives us information about aerosols, which is a very important element in our climate, both from global warming climate change perspective as well as pollution perspective. Next, we go to the thermal imaging system or TIR imaging, thermal infrared imaging systems or TIR systems.

The primary function of TIR imaging is to determine the temperature of the surfaces. That is the temperature of cloud tops, land surface and sea surface. And hence this is very important in climatology. Thermal infrared systems detect wavelengths between 3 micrometers and 15 micrometers. And the corresponding photons therefore have lower energies than the photons in the visible and the near infrared regions. The photonic energies are 0.1 electron volts to 0.4 electron volts. Which means the traditional CCD and photodiodes cannot detect the TIR range electromagnetic emissions. We have seen that the traditional photodiodes and CCDs can go up to 3 micrometers maximum. Okay. So, here specialized types of photodiodes are used including a few like indium antimonide based photodiodes. So, again the semiconductor material is indium antimonide which can detect radiation up to 5 micrometers. So, some part of the TIR spectrum this can capture. Others like like mercury cadmium telluride mercury 0.2 percent cadmium 0.8 percent and telluride 1 percent. So this these two hg 0.2 cadmium 0.8 this is one and telluride te okay mercury cadmium telluride photodiode and mercury doped germanium photodiode germanium which is doped with mercury. These two can detect up to 15 micrometers.

So, these specialist photodiodes, mercury cadmium telluride and mercury doped germanium are used for TIR imaging up to 15 micrometers. However, for these to be sensitive enough, the detectors themselves have to be cryogenically cooled by either liquid nitrogen so that the temperature is around 77 kelvins which is the maximum temperature possible or liquid helium which is 30 kelvins. So helium is liquid under 30 kelvins. So these detectors are cryogenically cooled so that infrared radiation from the detector itself does not interfere with the TIR image photons that are coming from the earth. So that extra cooling system is required for these detectors to work in the TIR range.

It improves the sensitivity of the detector. Now the spatial resolution is the same as before. So the entire discussion of VNIR imaging also goes over to the spatial resolution for TIR imaging systems. For spectral resolution, TIR imaging itself does not require a very high spectral resolution. The wavelengths are most often split into two bands, a 3 to 5 micrometer region just at the beginning of the far infrared region and an 8 to 15 micrometer at the end of the detection range. And in between the 5 to 8 micrometer we are avoiding because there is a high water absorption spectra there. So atmosphere is strongly opaque in that region. So we have two bands 3 to 5 and 8 to 15. The 3 to 5 micrometer band is more sensitive to temperature variations. So usually smaller the wavelength better is the sensitivity of that band in detecting small changes in temperature of the emitting surface.

So, the 3 to 5 micrometer band is more sensitive to small changes in temperature of the emitting surface may be sea, land or clouds than the 8 to 15 micrometer band. So, that is why 3 to 5 micrometer band is often favored. However, the problem with the 3 to 5 micrometer band is it is also part of the tail end of the shortwave radiation near infrared region. Even though near infrared region basically ends at 3 micrometers, there is still some and small amount of shortwave radiation in the 3 to 5 micrometer band as well, which interferes with the thermal radiation coming from the earth. So, it is susceptible to contamination from the tail end of the reflected solar radiation. So the 3 to 5 micrometer band is usually used only at night when there is no shortwave radiation. The 8 to 5 micrometer band can be used throughout the day and night because there is no shortwave radiation in that band but it is less sensitive. Now let us quickly look at the theory a little bit in order to understand how these systems work. Assume we have a detector that fully detects all incident thermal radiation within a certain band.

Say lambda max and lambda min. It may be lambda max may be 3, lambda min may be 5 or lambda max may be 15, lambda min may be 8 and does not detect anything outside of this band. So, we have simple filters removing everything else. Then the total radiative power E dot detected by the sensor due to emission from a surface of area A. So, we have again a ratio of area A over a solid angle delta omega. So, this solid angle is what is being projected onto your detector, correct? So, the total energy emitted by the surface of area A over a solid angle delta omega which is incident on your detector and suppose this surface of the blackbody radiation temperature called the bright often called the brightness temperature is given by this expression here.

$$\dot{E} = 2hc^2 A\Delta\Omega \int_{\lambda_{min}}^{\lambda_{max}} \frac{d\lambda}{\lambda^5 \left(exp\left(\frac{hc}{\lambda k_B T_b} \right) \right)}$$

$\dot{E}$ is the total energy that this detector is detecting at the detection end. T_b is the blackbody radiation temperature of the emitting surface or the brightness temperature of the emitting surface. λ_{min} and λ_{max} is the band, minimum wavelength and maximum wavelength of the band. A is the area of the ground area from which this emission is coming to that individual detector. $\Delta\Omega$ is a solid angle subtended by this area on the detector, $2hc^2$ you already know, this is the expression, ok.

So, $\dot{E}$ is what we are measuring, T_b is what we want to know, ok. Lambda we know, $A\Delta\Omega$ we can get from the geometry of the system. So, we can use an algorithm to evaluate the brightness temperature T_b, ok. Now, important point is depending on how the band filters are, where these band filters are located, this expression will change with respect to the brightness temperature. So, the x-axis is the obtained brightness temperature, y-axis is the normalized power. And we have three filters. One is there is no filter at all, 0 to infinity.

So, lambda min is 0, lambda max is infinity. It is detecting everything which is of course not usual. This is the solid line.

Now, here there is a 1. So, we are basically at 275 Kelvin. We are normalizing the detected power for any bandwidth by the actual detected power at 275 Kelvin and comparing that power with the power obtained at other brightness temperatures for that same bandwidth. So, for example, this is the unfiltered power obtained by your detector. Whatever that value is, suppose this is 5 watts is coming at 275 kelvins. Suppose at 288 kelvins, it is 5.5 watts with for that unfiltered band. then the value here is 5.5 by 1, 5. So, basically 1.1. So, 1.1 is the y-axis value and the x-axis value is 280 or 285. So, that is how this is being measured. Similarly, we are looking at two different IR bands, one varying from 10 micrometers to 11 micrometers. So, the lambda min is 10, lambda max is 11 micrometers.

Another band is 11 micrometers, 12 micrometers. So, lambda min is 11, lambda max is 12. This is the 10 to 11, this is 11 to 12. So, what you can see and all of these are again normalized with the actual value obtained at 275 Kelvin for that filtered band, and what you see is these lines have different slopes. So, depending on where you place your band, the E dot to Tb relationship is changing that the Tb to the power n if you say that n value is changing depending on where you are putting your band. So, for each definite band you need to have this correlation curve in order to find what is the brightness temperature that you would be obtaining for that band. So, using that say for example, suppose it is measuring say 0.5 watt per meter square with a band of 11 to 12 at 275 kelvins, it is now measuring 0.8 watt per meter, say 0.6 watt per meter square at 275 kelvins. Another different value of temperature then 0.6 by 0.5 basically 6 by 5 1.2 is the ratio so you go to 1.2 and this is 11 to 12 band so this is the dotted line you go down and the expected brightness temperature is measuring is around 287 Kelvin so using this curve we can based on the actual power obtained at the detector end, we can evaluate the blackbody temperature of the observed blackbody or brightness temperature of radiation. Now, there is one important caveat here that we will be discussing now. The quantity sensed at the detector is the brightness temperature of the incoming radiation. But what is actually needed is the brightness temperature of the emitting surface and there is a difference between these two because in the infrared region even in relatively transparent bands there is still significant atmospheric absorption and emission. Because of that attenuation and emission through the atmospheric interference, the observed brightness temperature and the actual blackbody surface temperature will be different from each other. So that effect needs to be corrected for. And there are two methods for doing this.\

$$T_{b0} = a_0 + a_1 T_{b1} + a_2 T_{b2}$$

We will discuss both of them. The first is the split window technique. Here, the brightness temperatures T_{b1} and T_{b2} are measured over two different but closely spaced spectral bands. So, we are measuring the same razor between two different and closely spaced spectral bands. One say between 10 to 11, another say between 11 to 12. Each of these is giving its own brightness temperature value for that certain location, razor.

Suppose the actual temperature of the surface is T_{b0}, then it can be shown that Tb0 will be a linear combination of T_{b1} and T_{b2}, where a_0, a_1 and a_2 are constants that can be separately evaluated through empirical means. Okay. So, these values are obtained from the physics of how the atmosphere interacts with the radiation in this wavelength region. So, given two bands, we can evaluate what this a_0, a_1 and a_2 would be for a given surface and the given atmospheric conditions, and these are already hard coded into your software. So, based on the actually observed T_{b1} and T_{b2}, you can get T_{b0}. Okay. The values, however, do differ between day and night time due to contribution of reflected sky radiation. Otherwise, the values are relatively constant for two given closely spaced bands as long as the emissivity of the surface is constant over these bands. This is true for the oceans. Oceans have a nearly gray emitter. So its emissivity does not change with wavelength as much, and so this technique is widely used for sea surface temperature evaluation.

Land surfaces, however, the emissivity does change with frequency, time of day, seasons, et cetera, vegetation covered. So this is more difficult. But with a combination of VNIR imaging, which will tell you what type of vegetation or land type that land system has, we do have sophisticated emissivity correlators that will give you the emissivity of the land surface and based on that, through a lookup table, we can get a_0, a_1 and a_2 values for various land surfaces as well. We can identify that specific land surface type, evaluate the values of a_0, a_1 and a_2 for that land surface type and then use the brightness temperature values to get the actual brightness temperature of land at that point. And the accuracy can be increased if we scan over multiple closely spaced wavelength bands so here closely spaced wavelength bands of say 0.1 or 1 nanometer is very useful and a linear combination of multiple brightness temperatures can be used to increase the measurement accuracy so that is one method the other method is called the two-loop technique so let us discuss how this looks like.

Basically once the sensor is measuring directly vertical the nadir view of the brightness temperature of land another time the sensor is measuring at an inclined with the 90 degree so there is a at an inclination of theta with respect to zenith so this can be done using a mirror apparatus for example okay so once a direct normal viewing and once an oblique viewing of the same result using at a different inclination. Now, the infrared radiation is moving upwards vertically here and at an incline here. So, it is passing through a greater depth of the atmosphere. This length is greater than this length by the theta value. So, this h and this z, hcos theta is z basically right so because of that the effect of the atmosphere is greater so the brightness temperature obtained here will be different than the brightness temperature obtained for nadir view okay. So, then we can use the typical atmospheric attenuation expression that we derived earlier in the class when we are looking at the how Infrared emission is passing up and down the atmosphere. We evaluated using the Beer Lambert's law of radiation absorption and emission. The expressions and we can use those expressions to derive these expressions as well.

$$T_{b1} = T_{b0}\exp(-\tau)+T_a(1-\exp(-\tau))$$

$$T_b = T_{b0}\exp(-\tau sec\theta)+T_a(1-\exp(-\tau sec\theta))$$

Where the brightness temperature for Nadir view is the actual land brightness temperature exponential of minus tau where the tau is the optical depth of the atmosphere for that frequency band plus the mean temperature of the absorbing layer of the atmosphere into 1 minus exponential minus tau. This expression T_{b0} is the ground brightness temperature, T_{b1} is the observed temperature, T_a is the mean atmospheric absorption layer temperature. This is for nadir viewing, for the inclined viewing this is the brightness temperature obtained $T_{b0}\exp(-\tau sec\theta)$ tau by cos theta because the height has increased. The distance has increased basically.

Here also minus tau sec theta is coming. Okay. Theta is known for the given frequency or wavelength value tau will be known. The mean atmospheric absorbing temperature can also be evaluated. Using this then we can find T_{b0} by relating these two expressions. Okay. So based on this then we can the two-loop technique will give you the ground brightness temperature alright and this can also be for oceans as well okay now how do we get the brightness temperature from the power okay so here what we have gotten. This is one method. Another method is through the radiation intensity value at a given wavelength. So, for a black body, the radiation intensity I lambda at a given wavelength is given by this expression here. We have known this. It is watt per meter square star radian, this expression. However, over an entire band, and this is the important part, using this expression here, we can get a different relation.

$$I_\lambda = \frac{2hc^2}{\lambda^5\left(exp\left(\frac{hc}{\lambda k_B T_b}\right)-1\right)}$$

$$(I_\lambda) = \frac{K_1}{\left(e^{\frac{K_2}{T_b}}-1\right)}$$

So, a certain band is set for a detector, say it may be 8 to 15 micrometer. The average radiation intensity I lambda over this band is being observed by our detector and this will be equal to certain instrument constants K_1 by $e^{\frac{K_2}{T_b}}$ this is the observed brightness temperature minus 1 so if K_1 and K_2 are given for the instrument for that specific wavelength so using that we can easily get the brightness temperature without having to integrate do a complicated integration so this integration is done for us and many instruments will give you the k_1 and k_2 values So, the brightness temperature observed is K_2 by log K_1 by the average radiation intensity plus 1. So, from at sensor brightness temperature obtained by this equation, the actual blackbody temperature of the surface can be determined using the atmospheric correction methods described up. So, first T_b is observed, T_{b1} and T_{b2}, using T_{b1} and T_{b2} and using the mean atmospheric temperature, this expression is evaluated, then from that T_{b0} is evaluated.

$$T_b = \frac{K_2}{\left(\frac{K_1}{(I_\lambda)} + 1\right)}$$

The emissivity of sea water is relatively constant at epsilon of 0.993. So this is easy to evaluate the sea surface temperature using this method. The land surface is somewhat harder because emissivity varies both spatially with location and temporarily time of day. But there are sophisticated algorithms nowadays so that land surface temperature can also be estimated. And this we can see here is a data from a satellite called Sentinel-3A for the month of May over the Asian, European and African continents. And this is the land surface temperature as obtained by that satellite through multiple runs over the May. So this is an average daytime land surface temperature. And you can see high temperature of 318 to 329 kelvins in western India, the Arabian desert and Sahara, whereas lower temperatures in southern African regions and Russia and northern Europe. So this way, a satellite will give us a global view of the land surface temperature using the TIR imaging systems. So, we will stop here.

We will finish TIR imaging in the next class and look at the other types of imaging sounding techniques we will begin in the next class. Thank you for listening and see you in the next class.